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MAGMAS AND IGNEOUS ORE DEPOSITS.

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PART II.

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ORE DEPOSITS FORMED BY MAGMATIC DIFFERENTIATION.

These deposits may be subdivided genetically as follows:

I. Formations due to crystallization—differentiation.

(*a*) In the *proto direction,* with an extensive enrichment of mineral crystallizing in the parent magma at an early stage.

(*b*) In the *restmagma direction,* with an extensive enrichment of elements or compounds which do not, or only to a subordinate extent, enter into the first crystallizing minerals.

II. Formations due to segregations of substances, originally dissolved in the silicate magma and precipitated in the liquid phase by refrigeration.

Some thirty years [34] ago I demonstrated that the deposits of titanic iron ore in gabbro, norite, etc., of chromite in peridotite, and of the platinum metals likewise in peridotite, were formed by a magmatic enrichment of the first crystallizing minerals, and also that the deposits of nickeliferous pyrrhotite were due to a process of magmatic segregation.

Now J. E. Spurr in his paper on " The Origin of Metallic Concentrations by Magmation " [35] and in his publication " The Ore Magmas " (1923) has taught us that we may formulate a class I*b* also, (see below).

I*a*. *Extreme Crystallization Deposits—Differentiation in the Proto-enriched Direction.*

As the most instructive examples, the segregations of titanic iron ore in gabbro and norite, and of chromite in peridotite (and pyroxenite) may be chosen. As demonstrated in previous papers [36] by other investigators and by myself—we often meet with " basic segregations " of " titanomagnetite "—gabbro and ilmenite-gabbro, or -norite, rich in ilmenite-magnetite, diallage, hypersthene, olivine, etc., *i.e.*, the minerals belonging to the first stages of crystallization of the mother rock. With a further advance in differentiation, the plagioclase disappears, and segregations of magnetite-ilmenite, accompanied by more or less diallage, hypersthene and olivine are formed. These segregations often carry a little spinel also. Finally there are segregations consisting almost exclusively of magnetite-ilmenite, that is, of the mineral belonging to a very early period of crystallization. These magnetite-ilmenite " rocks " often carry a fairly high percentage of spinel (such as 7, 10 or 15 per cent.), that is, of the mineral, which is the very first to crystallize in the intermediate stages of segregation. Apatite in these " rocks " sometimes is present in rather great quantity and sometimes is almost entirely

[34] See especially *Zeits. f. prakt. Geol.*, 1893, as to the deposits of titanic iron ore also previous Norwegian papers (1887 and 1891–92).

[35] This journal for 1923.

[36] See the literature list in Beyschlag-Krusch-Vogt textbook on ore deposits, I., sec. edit., 1914.

absent. This may depend on a separation of the titanic iron ores and the apatite with great differences in density and size of individuals, through the subsidence and the resorption of the minerals.

The formation of these ore deposits must be due to a pure gravitative differentiation (with accompanying resorption), without any direct coöperation of the volatile compounds. Thus the ore deposits may have crystallized from practically "dry" magmas. And in accordance therewith, pneumatolytic minerals are almost entirely lacking.

In the acid igneous rocks are geologically corresponding deposits of iron ores, carrying only a trifle of titanic acid, but often very rich in apatite. Examples are the gigantic deposits of Gellivare and Kiirunavaara in Northern Sweden and the small deposits at Solberg and Lyngroft in Southern Norway.

Ib. *Deposits Representing the Ultimate Residual Magma.*

In his above quoted papers Spurr gives, in my opinion, convincing arguments in explanation of the "veindikes" of gold-quartz of the "Yukon-Silver Peak type as an ultimate magmatic residual product."

We shall discuss some of his observations and arguments.

1. There exist continual transitions between granite, alaskite (granite very rich in quartz), arizonite (alaskite with the feldspars so subordinate that quartz is by far the most important constituent), granite pegmatite and dikes or veins (veindikes) of auriferous quartz.

2. These veindikes are limited to deeply eroded rocks and thus were formed at high pressure. They occur as lenses sometimes tabular, up to 50 or more feet in thickness and several hundred feet in length—and at Passagem in Brazil a vein has been followed even 6,000 feet in vertical depth or 10,000 feet on the incline. The veins or veindikes are not filling preëxisting cavities or openings. "The intrusions made their own room; they shoved and shouldered their own way in, pushing aside the shaly rocks, selected for invasion because the weakest of the sedi-

mentary series."[37] This proves it to have been a magmatic solution, of an igneous or aqueo-igneous character—and not a gaseous intrusion.

3. The accompanying minerals, such as tourmaline (sometimes present in great quantities), muscovite, hornblende, and other silicates, ilmenite, magnetite, pyrrhotite, etc., point to a rather high temperature.

4. Angular fragments of the wall rocks which do not touch either each other or the wall of the dikes sometimes occur in the veindikes—in the same manner as often in dikes of diabase. These fragments, in a structural sense differ from the brecciated inclusions of the hydrothermal fissure veins, and indicate about the same degree of viscosity as that characterizing the ordinary magmas, and thus a much higher viscosity (and density) than that of the hydrothermal solutions.

5. The quartz is as a rule massive, without banded or drusy structure, proving a high content of SiO_2 in the solution.

6. The metasomatic alteration of the wall-rocks differs as a rule rather strongly from that characterizing the gold-quartz veins formed at a rather shallow depth and due to the infiltration of gigantic quantities of (alkaline) water into the wall rock.

Concerning this "high temperature" and deep-seated group of gold-quartz veins Lindgren in his "Mineral Deposits"[38] remarks that these gold-bearing veins (of the Southern Appalachians) "appear to be connected with granitic intrusions, representing the final product of the most volatile part of the magma" —and this was in the main also my own idea until I had read Spurr's lately published papers. In view of his observations and arguments and many observations of the intimate connection between granite pegmatite and masses (dikes, etc.) of quartz and further the physicochemical explanations elborated principally by Jacob (1919) we may conclude that the main material of the "high temperature" gold-quartz veins has never passed a vola-

[37] This may be applied word for word to the "intrusive" pyrite deposits of Norway also (Sulitjelma, etc.). See Beyschlag-Krusch-Vogt, textbook on ore deposits, II., sec. edit., 1914, p. 340 et seq.

[38] McGraw-Hill Co., 1919, p. 676.

tile phase, but that it went directly over from the ordinary magma containing some H_2O, to an "aqueo-igneous" magma, illustrated by the formula $H_2SiO_3 \cdot nH_2O$.

Veins or veindikes often carry an admixture of the borosilicate tourmaline besides ordinary silicates, also magnetite, ilmenite, some carbonates, sulphides and arsenides, the latter sometimes in great quantities, and exceptionally also scheelite. On the other hand, minerals carrying tin, beryllium, lithium, or those carrying nickel and chromium, are practically always absent. And fluorine is present only in small quantity, principally entering into tourmaline. Minerals containing chlorine seem to be entirely wanting.

Thus the "aqueo-igneous" magmatic end-product may have been enriched principally in silica and water, but also in many cases in divers volatile compounds (carrying S, As, CO_2, etc.). There is, however, a remarkable difference between this mineral association and that characterizing the many subclasses of the Sn-W-Mo-Be-Li-B-F group of veins.

There is no distinct mineralogical or petrographical boundary between these gold-quartz veins (or veindikes), originally formed at a great depth and solidified at a rather high temperature, and the gold-quartz veins, crystallized at a much lower temperature. The material of the last named veins, that also occur in intimate connection with igneous rocks, thus may have passed a stage identical with the "aqueo-igneous" solutions of the Yukon and Silver Peak type. In the cooling of these solutions a great part of silica must have separated out, the solutions thus gradually passing over from an "aqueo-igneous" to a hydrothermal character.

To quote Spurr:[39] "Gold has not been concentrated from basic magmas." On the contrary, the auriferous quartz veins in general—irrespective of the depth at which they formed—are intimately associated with acid or intermediate igneous rocks (granite, quartz monzonites, quartz-bearing monzonites, diorites, gabbros, etc.) or co-magmatic provinces, that furnish, at the last

[39] "The Ore Magmas," 1923, I., p. 38.

stage, rocks of these compositions. All these magmas, at a late stage of the solidifications, may give rise to an acid (anchi-eutectic) restmagma, also under certain conditions to "aqueo-igneous" solution, illustrated by the formula $H_2SiO_3 \cdot nH_2O$.

The original parent magmas, from which the igneous rocks have been derived, carried a very small content of gold, by my theoretical working method [40] estimated to be 0.000,000,1 per cent. (= 1 mgr. per ton rock) and by Clarke and Washington [41] (1922) estimated to be 0.000,000,*n* per cent. (= 1–9, thus say 5 mg. per ton rock = 0.000,000,5 per cent.). These figures may give only an approximative idea of the true value.

Only a quite subordinate part of the original magmatic gold entered into the sulphides, precipitated or crystallized during the magmatic differentiation.[42]

In the ores formed by gases escaping at the various stages of crystallization of the magmas, gold is almost always entirely lacking, and when present usually only in a minute quantity.

From these and other observations, to be treated below, it may be concluded that the main quantity of the gold, dissolved as native metal in the magma, was enriched in the residual acid magma by the gravitative differentiation and from this again in the "aqueo-igneous" solution that consisted principally of $H_2SiO_3 \cdot nH_2O$. This may be why gold is usually associated with quartz.

The workable (!) auriferous quartz veins carry about 10–20 gr. gold per ton, equalling 0.001–0.002 per cent., an extremely low percentage according to ordinary standards. Under exceptionally favorable conditions quartz veins with only about 5 gr.

[40] *Zeits. f. prakt. Geol.*, 1898, pp. 324 and 388–390, Beyschlag-Krusch-Vogt, texthook on ore deposits, I., sec. edit., 1914, pp. 152, 157, 167–168.

[41] *Proc. Nat. Acad. of Sc.*, 8, May, 1922, and *Geophysical Laboratory*, Carnegie Inst., Washington, No. 462.

[42] The nickeliferous pyrrhotite, with about 3 or 4 per cent. nickel, carries on the average 1 part of gold to about 50,000 or 100,000 parts of nickel, equalling about 0.000,05 per cent. or 500 mgr. gold per ton ore. The pyrite of the "intrusive pyrite-group" (Sulitjelma, Røros, Huelva) carry on the average 2–3 per cent. copper, 25–50 gr. silver and about 500 mgr. gold per ton ore. And the blister copper from the copper ore from Røros and Sulitjelma contains about 0.05 per cent. silver and 0.0004 per cent. gold, or about 0.00002 per cent. gold, or 200 mgr. gold per ton copper ore.

per ton, equalling 0.0005 per cent. have been worked with profit. By far the most veins, however, carry still less gold. I might mention here that about 25 years ago my assistant in the metallurgical laboratory of the University of Kristiania, made a lot of analyses of quartz veins from divers Norwegian localities (in the Archean and the dynamo-metamorphosed cambrian-silurian formations), and found in every case a little gold, viz., 0.5–1.5, mostly about 1 gr. per ton, 0.0001 per cent.

From this I concluded long ago that the quartz veins, intimately connected with the igneous processes, and especially those which are not completely barren of sulphides, are always auriferous, though only a few of the numerous quartz veins are so rich in gold as to be of economic interest.

II. *Magmatic Sulphide Deposits Precipitated from the Liquid Phase.*

1. The nickeliferous pyrrhotites have already been dealt with.[43] Here it may be stated only that the sulphides of gabbroidic (noritic)—sometimes also peridoditic—magmas, dominantly FeS, NiF and Cu FeS_2, were segregated when by refrigeration the limit of solubility of the sulphides was reached. The temperature at this point being higher than the melting point of the pyrrhotite (or the mixture of pyrrhotite with pentlandite, chalcopyrite, pyrite, etc.), the sulphides were segregated in the *liquid* phase. Subsiding within the silicate magma the sulphide drops might collect gradually and locally masses of molten sulphide, often mixed with molten silicate, might result. The sulphidic, extremely thin, fluid magmas were often injected into the neighboring rock. In this manner we may explain the rather frequently brecciated character of the deposits and further the "offset deposits" in gneiss, at some distance from the mother norite.

The nickeliferous pyrrhotite deposits practically never contain pneumatolytic minerals and their petrographical features indicate a crystallization from an almost "dry sulphidic magma."

[43] See my previous treatises (as far back as 1893) and my paper in this journal for 1923, pp. 307–353.

2. The "intrusive pyrite deposits," of the type of Sulitjelma, Röros, Rio Tinto, etc., must also be explained as molten sulphides, segregated at high pressure from basic (gabbrodoic) magmas and afterwards injected, principally along the lines of schistosity of the neighboring rocks. These deposits differ materially from deposits of nickeliferous pyrrhotite, chemically and petrographically and partly also geologically. Here as stated by Thorolf Vogt, it may be a sulphidic magma rich in H_2O.[44]

No doubt future investigations will reveal still more classes of ore deposits formed by magmatic segregation.

ORE DEPOSITS FORMED BY ESCAPE OF VOLATILE COMPOUNDS.

Here we shall discuss the genesis of (1) the contact-deposits, (2) the veins and impregnations of the Sn-W-Mo-Be-Li-B-F group and (3) the Norwegian apatite veins.

Contact Deposits.—As is well known, contact deposits occur principally as replacements within limestone, and their characteristic mineral association is garnet (predominantly andradite, $Ca_3Fe_2Si_3O_2$), augite (often a hedenbergitic diopside, rich in $CaFeSi_2O_6$), hornblende, ilvaïte, biotite, wollastonite, epidote, vesuvianite, calcite and subordinate quartz. Boron minerals, such as tourmaline, axinite ($HCa_3Al_2BSi_4O_{16}$) and even ludwigite ($Fe_3O_4 \cdot 3MgO \cdot B_2O_3$) have been found in some deposits, as a rule, however, only as mineralogical curiosities, and they are entirely or almost entirely lacking in most localities. Fluorite is sometimes present, but, like other minerals carrying fluorine, on the whole is rather scarce, occurring only in traces or entirely absent. And the chlorine-containing silicate, scapolite, is when present, as it sometimes is, almost without exception subordinate only.

The contact deposits are connected with different igneous rocks, granites, quartz-bearing and quartz free monzonites, syenites, diorites, etc., rarely also gabbros and norites.

In many deposits iron is practically the sole heavy metal, but

[44] See the special paper which Thorolf Vogt is publishing in the Geol. Survey of Norway.

in the great metalliferous province of Arizona, New Mexico, Utah, etc., copper is the important heavy metal. Further, some deposits are worked for zinc, and quite exceptionally for lead. The latter contain usually a little silver. Gold also is present, but, generally, only in traces. Arsenic and bismuth are often found.

On the other hand, chromium and nickel (with cobalt) are practically absent, except in a few instances where nickel (with cobalt) occurs as a subordinate admixture. Tin is practically always lacking, as is also tungsten. Molybdenum is often present, but as a rule, in an inconsiderable quantity. Titanic acid in most deposits occurs only in traces, and up to the present time only a single contact deposit[45] rich in ilmenite has been described.

Particularly noteworthy is a special subclass of the contact deposits, viz. the *iron ore deposits,* with many of the small deposits in the Kristiania district and the more important mines in the Banat (Vaskó or Moravisca, etc.) and at Elba[46] as typical representatives.

In all these deposits magnetite is the principal, if not the sole, ore mineral. Specularite is often entirely absent, is sometimes subordinate to magnetite and only quite exceptionally predominant.

To 100 parts of iron the ores often carry 1 or 2, more rarely 3 or 4 parts of manganese; generally only 0.05, 0.1 or 0.2 parts of sulphur; and as little as 0.005–0.010, though oftener about 0.010, 0.020 or 0.025 parts of phosphorus; only occasionally does the proportion of phosphorus rise a little higher. Since the limestones (or occasionally also other rocks) which have been replaced by the iron ore in every case must have carried a trace of phosphorus, we may conclude that the emanations which gave rise to the deposits, contained no phosphorus at all or only in an extremely small quantity.

According to smelter experience, ores from contact iron-ore

[45] Viz. at Cebolla, California, described by Singewald, this journal for 1912.

[46] With these deposits the mines of Arendal (Klodeberg, etc.) in Southern Norway and of the Persberg-Dannemora type in Central Sweden are identical chemically and mineralogically, and, in my opinion, also genetically.

deposits, are practically barren of the metals nickel, cobalt, chromium, tin, wolfram, molybdenum, copper, lead and silver. The amount of copper at most mines is less than 0.002 or 0.005 parts to 100 of iron. Antimony, arsenic and bismuth also as a rule are entirely absent, but the ores from a few mines carry about 0.05 parts arsenic to 100 of iron.[47]

In many mines zinc is practically entirely absent, in others, however, it is present—partly as zinc spinel and partly as zinc blende—in quantity varying from .1 to .5 parts to 100 of iron. Minerals carrying fluorine (fluorite, tourmaline, etc.), chlorine (scapolite) and boron (tourmaline, axinite, etc., in the Banate ludwigite) are sometimes found, but are on the whole rather rare, and in many of these contact deposits of iron ore they are completely or almost completely lacking.

Of the small, but scientifically instructive deposits of the Kristiania district many years ago[48] I made a calculation concerning the occurrence of the deposits in relation to the igneous rocks: 20 per cent. of the total number (104) occur within the Silurian sediments (the limestones, as pointed out by V. M. Goldschmidt,[49] 1911) *immediately at the contact* with the igneous rock (generally granite, and quartz-sodium-syenite or nordmarkite) and not less than 16 per cent. occur in *xenoliths* (limestone), within the igneous rocks. These xenoliths are often of very small dimensions (such as 15–30 m. in length). The igneous rocks themselves show no magnetite in fissures and cracks, the iron ore being deposited only in the xenolites.

Principally on account of these observations I concluded in the above quoted papers, first, that the material of these iron ores, formed at a very high temperature, was derived from the igneous intrusion; second, that the material originally existed in a gaseous

[47] At some of the small contact deposits of the Kristiania district there is more bismuthite. At some blast furnaces running on these ores the incrustations of zinc oxide deposited on the upper part of the furnaces and in the flues, contain a small amount of bismuth oxide.

[48] Norwegian Geological Survey, No. 6 (*Geol. Føren. Førh.*, 1892), see also a previous paper in *Nyt Magazin for Naturvidenskaberne*, Kristiania, 9, 1884.

[49] "Die Kontaktmetamorphose im Kristianiagebiet," *Kristiania Vidensskabsselskabs Skrifter*, 1911.

condition, and third, that the gases escaped from the magma around the xenolites while it was, in the main, still molten.

The magmas of the granites and the other anchi-eutectic rocks, never being superheated, may perhaps have just begun to crystallize at the stage when the iron-bearing gases escaped from the magma. The result may be stated as follows: *the escape of the iron-bearing gases took place either just at the beginning, or at a very early stage, of the solidification of the magma.*

This conclusion regarding the period of formation of the iron ore deposits proper—carrying some manganese, but only a small admixture of the other heavy metals—with especial reference to the Kristiania district may, however, not apply to the deposits carrying principally chalcopyrite, zincblende, galena, etc. The magnetite deposits are sometimes traversed by granitic dikes, while veins of copper ore are sometimes clearly younger than granitic dikes. V. M. Goldschmidt, in his excellent paper on the Kristiania district, points out that the *exhalations of the compounds of copper, lead and zinc are younger than those of iron.* G. Berg [50] gives a lot of examples, from divers localities of contact deposits, of the comparatively late age of the sulphide ores, and he divides the pneumatolytic contact metamorphism into two stages.

A *first* stage of the formation of the iron silicates (andradite, hedenbergite, iron-rich diopside, further magnetite, specularite, bismutite and pyrrhotite partly) and a *second* stage of transformation of the iron silicates to amphibole and epidote, of formations of albite, and zincblende, galena, chalcopyrite, pyrrhotite partly and pyrite.

Veins and Impregnations of the Tin Group.—As is well known the cassiterite veins and impregnations, with Saxony, Cornwall and Mount Bischoff in Tasmania as typical representatives, are connected with granites (or sometimes quartz porphyry, or rhyolite) and especially with rather acid, alkaline granites (containing about 72–75 per cent. silica). The tinstone is accom-

[50] "On the Various Periods of the Pneumatolytic Contact Metamorphism," *Zeit. f. prakt. Geol.*, 1920, pp. 123–125.

panied principally by quartz, fluorite and other *fluorine*-bearing minerals, tourmaline and other *boron*-bearing minerals, divers lithium micas and other *lithium*-bearing minerals, topaz, beryl and other *beryllium* (glycium)-bearing minerals and *apatite* (fluorine apatite),[51] occasionally also other phosphates. The cassiterite deposits proper carry as a rule some wolframite (and scheelite), molybdenite, arsenopyrite, chalcopyrite, pyrite and pyrrhotite (both on the whole subordinate), often some bismutite, and as mineralogical curiosities, minerals of uranium, niobium and tantalum. Galena and zinc blende are commonly present, in most cases, however, only in very small quantities. Specularite, magnetite and ilmenite (with other titanium minerals) while sometimes present are only subordinate. And it is a characteristic feature that iron, the heavy metal incomparably the most widely distributed in the earth's crust, in this class of deposits recedes into insignificance.

The element association is thus Si, Sn, F, B, Be, Li, W, Mo, Bi, Cu, As, S, sometimes U, Nb, Ta, Sb and Pb, Zn, Ag;[52]—Fe, Mn, strikingly scarce; Ti almost absent, and Cr, Ni, Co practically entirely so; Al, K, and Na[53] are relatively subordinate; O of course in the oxides.

We point out especially the often astonishingly high quantities of fluorine and boron, both entering principally into tourmaline.

Taking the Cornwall district as an example J. H. Collins[54] has calculated—or in part estimated—the following quantities per yard of depth:

About 21,250 tons of metallic tin,
About 2,550,000 tons of fluorine,
About 2,500,000 tons of boron,

or *more than 100 times as much fluorine (and boron) as tin.*

[51] Sometimes with a very low percentage of chlorine.

[52] As to the Bolivian tin-silver veins I refer to some remarks in the next chapter.

[53] According to Tronquoy (*Bull. Soc. Franç. de Min.*, 1912, 35) besides tourmalinization, a great albitization, has taken place in the cassiterite deposits of France. Here then, we meet with a rather considerable addition of potassium. Concerning aluminium, see below.

[54] *Journ. of the Royal Institute of Cornwall*, XXXIX., 1893, pp. 160–163.

In some veins the amount of cassiterite decreases or entirely disappears, and other minerals predominate in its place. We may therefore formulate a great Sn-Wo-Mo-Be-Li-B-F group, made up of the following types of deposits.[55]

Predominant cassiterite,
Cassiterite and wolframite,
Wolframite without cassiterite,[56]
Molybdenite and wolframite without cassiterite,
Molybdenite and bismutite,
Molybdenite and copper ores,
Molybdenite with neither wolframite nor cassiterite and with only an inconsiderable admixture of chalcopyrite, pyrite, etc.,[57]
Divers lithium-minerals, especially lithium phosphates (amblygonite, $(Li\ F)_2 \cdot Al_2O_3 \cdot P_2O_5$) in some veins with a subordinate admixture of cassiterite,
Cryolite (Ivigut, Greenland, accompanied by cassiterite, wolframite, "lithium-cryolite," etc., a little topaz but no boron minerals),
Beryl (smaragd) and topaz (*e.g.*, Minne at Eidsvold, Norway),
Cassiterite and copper sulphide ores in about equal proportions.
These grade into veins of sulphide ores, principally copper ores, that mineralogically and geologically differ so much from the cassiterite veins proper, that I prefer to treat them separately.

These deposits of cassiterite, wolframite, molybdenite, etc., occur as veins or impregnation either within the granite itself, or in the adjoining country rocks. In Cornwall the veins of cassiterite have been mined to a depth of 3,000 feet, proving that this granite, at the period of vein-filling, was solidified to a considerable depth. At some other localities, as in Zinnwald, Saxony, the flat lying cassiterite veins, following the joint planes of the granite (dipping 15–30°) occur only in the *upper* part of the granite batholith. Further the veins are sometimes cut by younger dikes of a fine-grained granite. Consequently, the Saxon geologists[58] have concluded that the formation of the

[55] I refer to the details in Beyschlag-Krusch-Vogt, text-book on ore deposits, part II., sec. edition, 1914, p. 507 et seq.

[56] Wolframite further occurs in some hydrothermal copper-lead-silver veins.

[57] Such veins, besides quartz, also carry some tourmaline. They occur in great number connected with granite in Norway, where deposits of cassiterite are entirely lacking.

[58] R. Beck, *Zeits. f. prakt. Geol.*, 1896, p. 149, and many papers of Dalmer and others.

cassiterite deposits here took place at a period when the upper part of the granite batholith had solidified, but *before the solidification of the deeper part.*

As is well known, the mineralization has produced a briefly metamorphosed greisen zone as a rule rather rich in boron and fluorine minerals, as well as ore minerals. Within the greisen pseudomorphs of cassiterite or tourmaline after feldspar, and of cassiterite and topaz after quartz, establish the fact that the granite enclosing the veins was entirely solidified at the time of their formation. In some districts, (*e.g.,* Pitkäranta, Finland), where the granite is accompanied by contact deposits as well as by cassiterite deposits, the latter are younger than the former.

Norwegian Apatite Veins.—The Norwegian apatite veins,[59] bear the same relationship to gabbro as the cassiterite veins do to granite. These veins occur in several Archean districts of Norway and in one district also of Sweden and are without exception, intimately connected with gabbro (especial olivine gabbro, with about 46–48 per cent. silica). The mineral association is as follows:

Apatite (chlorine apatite) and, as a mineralogical curiosity, also wagnerite (Mg_2FPO_4);

Magnesian mica, especially phogopite, and enstatite; further, diopsidic pyroxene, hornblende, scapolite (chlorine-bearing), an acid plagioclase (rather rare), quartz (in the veins within the gabbro itself almost completely absent), calcspar (only in some few veins);

Divers titanium minerals, especially rutile (sometimes predominant), ilmenite and titanite, further also pseudobrookite (rare);

Specularite (sometimes in quantity), magnetite, pyrrhotite (sometimes in quantity) and pyrite;

As mineralogical curiosities chalcopyrite and divers other minerals;

As minerals of younger date, prehnite, etc.;

Tourmaline has been found in a few veins only.

Fluorite is almost completely missing, and the apatite from these Norwegian deposits is a typical chlorine-apatite. On the

[59] See Brøgger and Reusch, *Zeits. d. Deutsch. geol. Ges.,* 1875, and *Nyt Magazin for Naturvideskaberne,* 25, 1879, and a paper by myself in *Zeits. f. prakt. Geol.,* 1895, pp. 367–370, 444–459, 465–477, further Beyschlag-Krusch-Vogt, text-book on ore deposits, in a chapter following the chapter on the tin deposits.

other hand, the apatite from the Canadian veins that are in the main analogous carries predominant fluorine as well as chlorine.

The total production of apatite from the Norwegian veins amounts to about 150,000 tons, by far the most from Ødegaarden in Bamle.

Some deposits carry predominant rutile (and have been worked in a small scale for this mineral); others are rather rich in iron ore, etc.

The apatite veins traverse the gabbro, especially along joint planes and they are themselves traversed by diabase dikes, derived from the gabbrodic magma. For this and other reasons the apatite veins—like the cassiterite veins—must belong to a late stage of the solidification of the mother rock. And the association of the minerals, often occurring in remarkably big individuals, indicates formation at a rather high temperature.

With the apatite veins also, there is an intensive metamorphism of the wall-rock, the gabbro being transformed into the well-known scapolite-hornblende rock; the diallage and olivine to hornblende (or sometimes bronzite), and the plagioclase to scapolite. This mineral may be considered chemically as a plagioclase with an admixture of NaCl, resp. CaO; the meionite silicate of the scapolite theoretically carrying 93.75 per cent. An-silicate, $CaAl_2Si_2O_8$, and 6.25 per cent. CaO—and the marialite-silicate 93.10 per cent. Ab-silicate, $NaAlSi_3O_8$, and 6.90 per cent. NaCl = 4.19 per cent. Cl. According to several investigations, the scapolite of the scapolite-hornblende rocks contains about 2.5 per cent. Cl and the rock itself about 1.5 per cent. Cl.

The width of the scapolite-hornblende rock formed on both sides of the veins, varies with the width of the vein: with veins 1 cm. thick, it will be some few cm. in width; with veins 10 to 20 cm. thick, from ½ to 2 meters, and at Ødegaarden, where the gabbro is traversed by numerous and often rather thick veins, by far the greatest part of the original gabbro is transformed to the new rock.

According to a calculation made by myself (*Zeits. f. prakt. Geol.*, 1895, pp. 458–459) the mass developed by mining at

Ødegaarden amounting to 5,800,000 cubic meters, contained 1.19 per cent. apatite (= 0.49 per cent. P_2O_5 or 0.21 per cent. P) in the veins, and about 1.4 per cent. Cl mostly in the scapolite-hornblende rock, but a small amount in the apatite. Although these figures may be open to some criticism, we may safely conclude that the *quantity of chlorine* in the traversing solutions was *many times that of phosphorus.*

The element association of the apatite veins Si, Ti, Cl, P, Fe, Mg, Ca, Na, etc., stands in the same chemical relation to the gabbro (or the gabbro magma) as the element association of the tinstone veins to the granite (or the granite or pegmatite magma).

General Discussion.

In view of the intimate connection of the cassiterite veins with granite, the considerable amount of fluorine, the analogy with the pneumatolytic formation of minerals at recent volcanoes, and Daubrée's epoch-making synthesis of cassiterite by the reaction between $SnCl_4$ and H_2O, Daubrée and Elie de Baumont, as far back as 40's of last century advanced the well-known theory that the minerals of the cassiterite deposits were formed by volatile fluorides, such as SnF_4, BF_3, etc. This theory some thirty years ago[60] I extended to a certain degree, explaining the fluorides as a "magmatic acid extraction," due to the reaction of HF, etc., on the granitic magma in a rather advanced stage of the solidification.

These cassiterite and apatite veins are younger than the surrounding mother rock; on the other hand, observed facts (see previously) indicate, that they were formed at a period when the deeper parts of the magma were still molten.

Having a very important bearing on this conclusion is the well-known fact, that many of the cassiterite minerals (cassiterite, wolframite, molybdenite, tourmaline, topaz, beryl, divers lithium-minerals, etc.) are also characteristic of the granite pegmatites. The material of these minerals, as above explained, may be due

[60] *Zeits. f. prakt. Geol.*, 1895, p. 473 et seq.

to a magmatic extraction, especially at a more advanced stage of the solidification of the magma.

The element association of the cassiterite veins is a function of the chemistry of the granite, and that of the apatite veins of the chemistry of gabbros. In both groups of deposits there is an *extraordinarily great quantity of a haloid,* in the one case *fluorine* and in the other case *chlorine.* Remembering further the great chemical activity of these haloids we may conclude that the " magmatic extraction " of these deposits may be due to the compounds of fluorine and chlorine.

The above theoretical explanation suggests that a notable extraction of tin from the magma is possible only through a relatively great quantity of HF to H_2O. This would explain the geological observation that a greater formation of SnF_4 takes place only in the *granitic restmagma.* Here also the magmatic contents of SnO_2 may have been concentrated.

The same may hold true also with boron, glucinium, molybdenum, tungsten, tantalum, niobium and uranium, whose average contents in the rocks are estimated to be: 0.001 per cent. B, 0.001 per cent. Be, 0.000*x* per cent. Mo, 0.00000*x* per cent. W, 0.00000*x* per cent. Ta, Nb or Y. In the granitic magmas in a rather far advanced stage of the crystallization we may presume higher figures.

That boron in the igneous rocks and allied mineral deposits, including those connected with basic magmas, enters, to a remarkably great extent, only into the pneumatolytic minerals (tourmaline, axinite, etc.) may be due in the main to the extraordinary low boiling points of its haloids. The gas tension of these compounds at the magmatic temperature may be exceptionally high.

Even in the granitic restmagma the percentage of the elements Mo, W, U and Nb, Ta must be inconsiderable. That only these elements are so considerably enriched relatively in the Sn-W-Mo-Be-Li-B-F veins may be connected with the fact that the high valence haloids of all these elements are characterized throughout by extraordinary low boiling points, the fluorides as low as 19.5–229°. The haloids of bismuth and arsenic are also easily volatile

And BeF_2 has a much lower boiling point than the fluorides (and chlorides) of the next related elements, Mg, Ca, Sr and Ba.

The question, why among the alkalies only lithium is so remarkably enriched in the tin stone-veins, seems difficult to answer. We shall only remark that within Mendeléef's vertical columns I., II., III., Li, Be and B have the lowest atomic weights (and F the lowest in the column VII.).

As to the phosphoric acid, abundant in the "chlorine"-veins of the olivine gabbro, (on the average about 0.35–0.45 per cent. P_2O_5 in the rock) and also very often present in the cassiterite-veins of the acid granites (with about 0.02–0.05 per cent. P_2O_5 in the rock) we call attention to the very low boiling point and critical point of PCl_3 and still more of PF_3.

Concerning the abundance of the titanium minerals in the apatite-veins, and their rarity in the tin-veins, note that TiO_2 is present in much greater quantity in the gabbros (olivine gabbros, often with about 1.5 per cent. TiO_2) than in the acid granites (on the average with only 0.1–0.25 per cent. TiO_2), and further the low boiling point of TiF_4 and the still lower one of $TiCl_4$.

As to the heavy metals, copper, lead, and zinc, (with cadmium), the melting points of the fluorides of these elements are much higher than the melting points of all the above mentioned elements, with the exception of lithium. Their gas tensions at a temperature of say about 800° are thus rather low. To still unknown factors may be due the scarcity of zinc and lead in the great Sn-W-Mo-Be-Li-B-F group compared with copper sulphides of which a special class stands rather close to the tinstone group.

The problems concerning magmatic extraction evade direct observation, and we may be satisfied if we can draw up the main lines.

An open question is why the mutual proportions between the characteristic elements (Sn, W, Mo, Be, P, etc.) varies so highly in veins that in other respects present much the same geological and mineralogical features. Suffice it to say that even small variations in the quantity and mutual proportion of the reacting

agents (H_2O, HF, HCl, CO_2, etc.) and variations in pressure and temperature may occasion a very pronounced alteration of the magmatic extraction.

The deposition of the minerals from the escaping gases may be illustrated by cassiterite.

A gas consisting of SnF_4, HF and H_2O must deposit cassiterite: (*a*) with decreasing pressure; (*b*) with decreasing temperature; and (*c*) by an alteration in the proportion between SnF_4, HF and H_2O, viz. by a relative decrease of HF and a relative increase of H_2O.

Each of these factors occasions the crystallization of cassiterite from the rising gases. A small decrease of the pressure or of the temperature probably exerts an inconsiderable influence,—but, on the other hand, even a small diminution of the percentage of HF in the gases may react very strongly. A relatively great quantity of HF, but only a small quantity of H_2O is consumed in the formation of the minerals of the greisen zone, and of the vein-filling (fluorine-bearing mica, tourmaline, topaz, etc.). Thus the rising gas may become continuously relatively poorer in HF, and this fact is probably the principal factor in the deposition of the cassiterite.

As is well known, at Cornwall there is a very striking *primary depth difference,* the upper primary zone carrying sulphide copper ores with only subordinate cassiterite, then follows a transition zone, while the still deeper zone carries predominant cassiterite, with subordinate copper pyrites, etc. Thus in the main the deposition of the cassiterite stopped at a certain elevation, while the deposition of the copper ore took place still higher up. The upper limit of the deposition of the cassiterite is probably determined by the three factors above mentioned.

The minerals within and immediately adjoining the contact deposits of the Kristiania district are characterized by larger size than those within the schists of normal contact metamorphism. Previously (1892) I designated this phenomenon "*potentiated* contact metamorphism," which might be explained by the great amount of H_2O, etc. in the magmatic gases penetrat-

ing the sediments where the ores were deposited. Influenced by the old classic French school, in my old treatises (1884, 1892) I put forth the hypothesis that the iron in the gases existed as chloride or fluoride, and Goldschmidt (1911) postulated gases such as FeF_3 and SiF_4 or $FeCl_3$ and $SiCl_4$.

Years ago, however, I reached the opinion that the gases depositing the contact ores were not, in the main, haloids. Fluorite, tourmaline, scapolite, etc., certainly do occur in iron contact deposits, proving that these minerals may form under the physical and chemical conditions present. But the small quantity of F, Cl (and B) in some contact deposits and their complete absence in others in my opinion may indicate that the haloids were not the mineralizers that produced the contact ores. The prominent difference between the element-association in the casserite-veins and the contact metamorphic deposits—in the one much Sn, W, Be, Li, P, in the other practically none of these elements; in the one remarkably little iron, in the other predominant iron—also indicates that the chemistry of the magmatic extraction in the two cases may have differed materially. Sometimes both classes of deposits occur with the same granite, which also suggests that the chemistry of the magmatic extraction of the two may have deviated widely. If then fluorine was the essential agency in the formation of the cassiterite veins, it must, from that fact be excluded from an essential rôle in the iron contact deposits.

Which agents were active in the magmatic extraction of the iron, etc. of the contact ores, is an open question. As a working hypothesis one may think of H_2O—this being abundant at the beginning of the solidification of the magma—together with CO_2. This might give $FeCO_3$ or $FeH_2(CO_3)_2$, soluble in H_2O,[61] a similar manganese compound and further a $H_2O + SiO_2$ compound. The traces of HF, HCl might yield small quantities of the extremely light volatile fluorides or chlorides of boron, bismuth, arsenic, etc.

At Orijärvi in Finland, according to P. Escola,[62] in the con-

[61] At supercritical temperatures there is no difference between the gaseous and the liquid phase.

[62] *Bull. Com. Geol. de Finlande*, No. 40, 1914.

tact aureole of an Archean oligoclase granite, a limestone has been supplied with not only silica and iron oxides, besides the ordinary sulphides of divers heavy metals (Cu, Pb, Zn), but also with a great quantity of *magnesia*.[63]

Th. Vogt (April 16, 1920) noted that the difference between

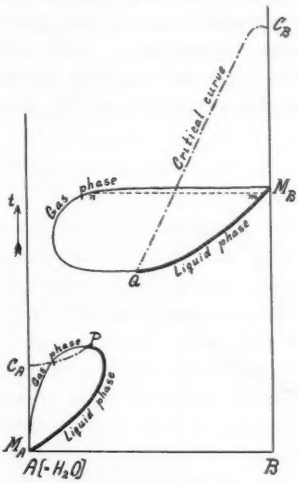


FIG. 4*a*. Contact iron ore deposits of the Kristiania type, at a rather shallow depth (and thus at a rather limited pressure), the gases escaping at the beginning of crystallization.

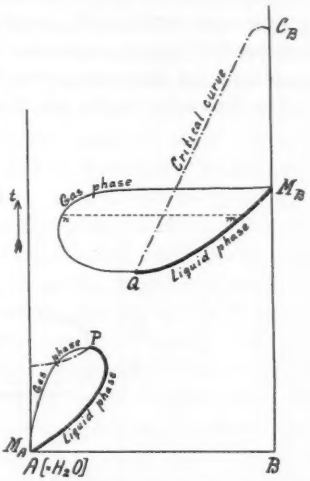


FIG. 4*b*. Contact iron ore deposits of the Orijärvi type, at a very considerable depth (and thus at a very high pressure), the gases escaping at a somewhat more advanced step of solidification. According to Th. Vogt.

the composition of the iron ores in the Kristiania district (iron oxides and silica, with an inconsiderable percentage of magnesia)

[63] A high percentage of MgO is found at many other iron deposits that in my opinion may be considered deep seated contact-deposits. Thus the iron ore from Klodeberg at Arendal contains on the average 0.50 MgO: 0.50 CaO, and the iron ore from Meløvær in the North of Norway on the average even 0.87 MgO: 0.13 CaO. Also many of the iron ore deposits in Central Sweden carry much MgO in relation to CaO (but only rather small quantities of Al_2O_3).

and the composition of those at Orijärvi (iron oxides and silica, with a high percentage of magnesia) might be connected with the depth of the igneous rocks,—those of Kristiania having solidified at a relatively shallow depth (say about 1–1.5 km.), those of Orijärvi on the contrary at a considerable depth. The increased pressure in the latter might prevent the escape of the volatile compounds until a more advanced stage of the crystallization, and thus at a somewhat lower temperature. Under such conditions the magmatic extraction might be somewhat altered.

The theoretical results are illustrated graphically in Fig. 4*a–c*.

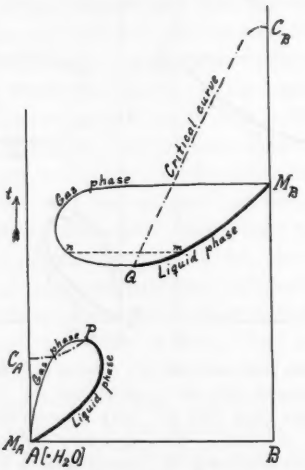


FIG. 4*c*. Cassiterite-veins, the gases escaping at a very advanced stage of the solidification of the granitic magma.

The volatile compounds depositing the contact iron ore deposits may have escaped from the granitic magmas at a temperature of about 1000° (or sometimes somewhat below 1000°)—those depositing the sulphidic ores at a somewhat lower temperature,—and those depositing the cassiterite-veins at a still lower temperature, say about 750°.

In their passage through the rocks of the contact aureole, or, with the cassiterite veins, through the upper already solidified part of the granite, or through the surrounding rocks, the gases may have been somewhat cooled, especially in the outer zones of the contact aureoles and in the surrounding rocks of the granite. The final deposition of the minerals may thus have taken place at a temperature several hundred degrees lower than the temperature of the gases at their escape from the magmas. The quartz of the cassiterite veins, according to present investigations, seems in most cases [64] to be a β-quartz, formed below 600–625 degrees.

These contact deposits of (oxide) iron ore show unbroken transitions to those carrying more or less pyrite, chalcopyrite, zinc blende, galena, etc., the latter ores in general being of a somewhat younger age. And the Sn-W-Mo-Be-Li-B-F veins, sometimes carrying much copper sulphide ore, continuously grade into veins of copper ores, carrying no tin, tungsten or molybdenum. Telemarken and neighboring districts in the South of Norway furnish examples. Here occur, intimately connected with very old granites, many veins and impregnations of molybdenite, some poor and some rich in copper sulphide ores, and further many veins of pure sulphide copper ores (primary chalcocite and bornite, often accompanied by some specularite; chalcopyrite sometimes alone and sometimes accompanied by a little bornite, but without pyrite or pyrrhotite, thus stoichiometrically with just as much copper as iron or somewhat more copper than iron). The gangue minerals of the copper veins are principally quartz and muscovite, sometimes subordinate tourmaline and fluorite. In the same district, in the neighborhood of chalcopyrite, and of molybdenite veins, occur a few large veins of fluorite, with only a trace of copper. The copper veins show muscovite grown on the wall rock, the structure of these veins being identical with those of Zinnwald, and the veins of the copper ores within granite or granite dikes are accompanied by a

[64] R. Beck (*Centralbl. f. Min.*, etc., 1912, 693–698) thinks that the cap-quartz from the cassiterite-veins of Geyer in Saxony is α-quartz (formed above 600–625°).

typical greisen-zone, here especially developed as a mica-rock. For these reasons many years ago [65] I designated these veins as "cassiterite-veins with sulphide copper ores instead of cassiterite." Further an area of quartz diorite at Svartdal in Telemarken contains quartz-tourmaline veins, extremely rich in tourmaline, carrying pyrite, chalcopyrite, considerable bismutite, and some gold.

Copper veins, rich in tourmaline besides quartz, connected with acid, intermediate, and sometimes with basic, igneous rocks, are represented on a great scale in many districts of Chile, and other countries. From Chile also have been described [66] quartz-tourmaline veins carrying cobalt ores (cobaltite) and connected with a basic igneous rock. Quartz-tourmaline veins carrying predominant galena and zinc blende are rather rare. All these ore deposits may be explained by the escape of gases during the solidification of the magmas, probably at a rather advanced stage of the solidification. The deepseated auriferous quartz veins often carry tourmaline, and other silicates. These veins may represent the residual $H_2SiO_3 \cdot nH_2O$-magma in which more or less of the volatile compounds have been concentrated.

(*To be concluded in No. 5, due Aug. 1.*)

[65] See my review in *Zeits. f. prakt. Geol.*, 1895, p. 149 et seq.
[66] Stutzer, *Zeits. f. prakt. Geol.*, 1906, 294–298.

DETERMINATION AND USE OF THICKNESSES OF INCOMPETENT BEDS IN OIL FIELD MAPPING AND GENERAL STRUCTURAL STUDIES.[1]

W. W. RUBEY.

THE thicknesses of rock units must be known in most structural and stratigraphic studies and in order to prepare structure contour maps. The value of the measurements depends almost entirely upon their accuracy yet the development of the fundamental concepts underlying the computation and use of thicknesses seems to have lagged behind the refinement of field methods. Stratigraphic thicknesses are generally determined and used on the assumption that rocks are strictly competent despite the fact that, in most regions where the relatively precise methods of the oil geologist are applied, very incompetent shale beds constitute a large part of the stratigraphic section.

This paper grew out of a study of some steep folds in thick shale series in northeastern Wyoming but the theoretical conclusions reached perhaps have a more general application. An attempt is made to demonstrate by a theoretical analysis (1) that thicknesses of incompetent strata may be determined by methods somewhat simpler and perhaps more accurate than those generally used, (2) that, in such rocks, the practice of "correcting for dip" in preparing structure contour maps is unwarranted, and (3) that neglected evidence bearing on the origin and nature of folding exists in the usual measurements of thickness. No claim is made that the suggestions here offered lead to final accuracy but the premises upon which they are based seem to be more in harmony with accepted geologic principles than the as-

[1] Published by permission of the Director, U. S. Geological Survey, Washington, D. C.

sumptions commonly made and the methods of their application are at least as simple as the others.

The writer gratefully acknowledges the helpful criticism received from many of his colleagues of the U. S. Geological Survey, especially D. F. Hewett, Frank Reeves, P. G. Nutting, M. R. Campbell, and K. C. Heald.

THE DEFORMATION OF ROCKS INTO SIMILAR FOLDS.

Assumptions.—The saying is trite but true, that nothing better is likely to come out of the mathematical machine than that which is put into it, but deduction from reasonable assumptions is worth while if it yields stimulating or useful conclusions or if it demonstrates that the original premises are not true. By assuming (1) that the density and volume of shale remains constant during folding, (2) that the dimensions are changed in only one plane, (3) that shale is sufficiently plastic to be folded into strictly similar folds, and (4) that the axial planes[2] of the folds are vertical, then several important characteristics of deformed shale beds can be determined.

1. In a mathematical analysis of any fold, whether of the parallel or similar type, an assumption of constant volume during deformation must be made. This assumption certainly is not strictly true where deformation has been intense and there has doubtless been some decrease of volume even in gently dipping beds, but, where the textural or mineralogic changes in the rock are negligible, the volumetric decrease due to normal consolidation probably has been fairly constant throughout the extent of the rock.

2. The assumption that beds are squeezed or stretched in only

[2] In its original sense the term *axial plane* refers to the plane bisecting the angle between the limbs of a fold (Willis, Bailey, "The Mechanics of Appalachian Structure," U. S. Geol. Survey, 13th Ann. Rept., Pt. 2, p. 220, 1892). This definition is not exact nor according to present usage for those folds in which one limb has been squeezed more than the other, as is the case in many drag folds. In this paper the term is used for the plane which roughly bisects the limbs of the fold, and connects the lines of maximum curvature on different strata. Strictly speaking, *plane* is not a precise term for axial "planes" may curve gently and some geologists prefer to use *axial surface*.

two dimensions is not entirely valid since it requires that folds be of infinite length and without minor axes, but in any fold with a well-defined major axis the error is slight and the mathematical simplicity of the assumption is convenient for purposes of discussion.

3. That incompetent shale beds are deformed into similar folds instead of into the parallel or concentric folds [3] characteristic of more competent rocks is an assumption that perhaps is generally accepted. This concept is based chiefly on the observed behavior of strata in small squeezed folds but it has been extended to include also larger and gentler folds in the more plastic rocks. The argillaceous rocks, though highly plastic, are of course not perfectly so and under light load or rapid deformation their behavior undoubtedly approaches that of more competent material and a complete gradation between similar and concentric folds is to be expected. However, most shales are relatively so plastic or incompetent that, even under moderate loads, they probably are deformed into folds more nearly similar than concentric.

4. Unlike the other assumptions, the probability that the axial planes of similar folds are vertical is not widely recognized but to the writer the assumption seems warranted. As will be discussed more fully in another part of this paper, essentially similar folds may be formed by stresses acting either horizontally (tangential compression) or vertically (uplift or depression). In general, the axial planes of similar folds due to horizontal compression would be expected to be vertical—that is, at right angles to the direction of thrust—but minor folds superposed upon a larger one might have their axial planes tilted in one direction if the broader warping was later or in the other direction by drag folding due to differential movement between competent beds. Under vertically acting forces the axial planes of similar folds would also tend to be vertical but with a lapse of time or during gradual deformation the resistance of the strata to tension would

[3] Van Hise, C. R., "Principles of North American pre-Cambrian Geology," U. S. Geol. Survey 16th Ann. Rept., pt. 1, pp. 599–600, 1896. The term *concentric* is preferred by some geologists; see Willis, Bailey, "Geologic Structures," pp. 19, 21, 154, 1923 and Leith, C. K., "Structural Geology," pp. 172–175, 1923.

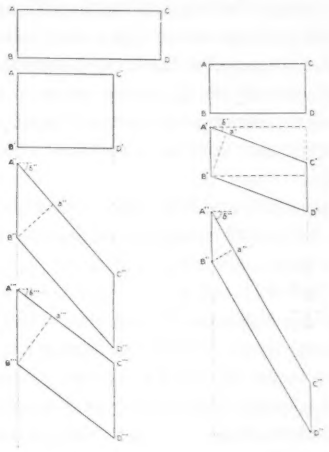
cause them to slump and the axial plane of an asymmetric anticline would probably incline in the same direction as the dip of the beds, that is, toward the steep side of the fold. This adjustment might also occur in the higher beds in asymmetric folds formed by horizontally acting forces. With our present knowledge, in many places the angle or even the existence of such inclinations can not be safely forecast, and it is therefore probably safer to assume that the axial planes are vertical.

Special Terms Used.—For the thickness of the strata where unmodified by deformation, *original thickness* is used. In concentric folds the geologist is concerned solely with the *stratigraphic thickness,* that is, the measurement of the strata made at right angles to the bedding, for this thickness remains the same as the *original thickness* during folding. In similar folds, however, thicknesses vary with the degree of deformation and position on the fold and the *vertical thickness* (the distance across beds measured in a vertical direction) is also important.[4] Where the strata are horizontal, the *original, stratigraphic,* and *vertical* thicknesses may be the same.

The Effect of Horizontal Thrust upon Plastic Rocks.—Under horizontal thrust or compression, rocks that are sufficiently plastic will be shortened and correspondingly thickened. During deformation the beds might remain horizontal (Fig. 1, $A'B'C'D'$) and be thickened uniformly throughout the area affected by the shortening and in this case the new modified thickness, $A'B'$, would be equal to the product of the original thickness, AB, and the ratio of shortening, $AC/A'C'$. This is simply a mathematical statement of the amount of the commonly recognized thickening on the crests and in the troughs of ideal similar folds. In most places, however, the beds would be tilted during the deformation (Fig. 1, $A''B''C''D''$ and $A'''B'''C'''D'''$) and this would result in more complicated thick-

[4] In an interesting mathematical analysis of some of the problems of folded strata, E. L. Ickes ("Similar, Parallel and Neutral Surface Types of Folding," pp. 576–580, ECON. GEOL., vol. 18, 1923) has called the thickness parallel to the axial plane the *ordinate thickness,* a term that is useful because it is applicable whether the axial planes are inclined or vertical.

ness relations. If the axial plane is vertical, the *vertical thickness* of any stratum in the similar fold would remain constant regardless of the amount of tilting (Fig. 1, $A'B' = A''B'' =$



1 2

FIG. 1. Deformation of incompetent material by horizontal compression. Areas $ABCD = A'B'C'D' = A''B''C''D'' = A'''B'''C'''D'''$. $A'B' = A''B'' = A'''B''' = AB \times AC/A'C'$. $a''B'' = A'B' \cos \delta'' = \cos \delta''\ AB \times AC/A'C'$. $AC/A'C' = A'B'/AB = \dfrac{a''B''}{\cos \delta''\ AB}$.

FIG. 2. Deformation of incompetent material by vertical uplift or depression. Areas $ABCD = A'B'C'D' = A''B''C''D''$. $AB = A'B' = A''B''$. $a'B' = AB \cos \delta'$.

$A'''B'''$, since the area of a parallelogram equals the product of its base and its altitude). That is, the thickness of a bed measured parallel to the axial plane would be dependent wholly upon the ratio of shortening and be quite irrespective of the amount of dip. However, the *stratigraphic thickness* would vary with both the ratio of shortening and the dip of the stratum,—the stratigraphic interval, $a'''B'''$, would be equal to the product of the *original thickness, AB,* the ratio of shortening, $AC/A'C'$, and the trigonometric function of the angle of dip, $\cos \delta'''$.

In a similar fold with a given ratio of shortening, the modified *stratigraphic thickness* is the same as the *original thickness* at but one angle of dip, for only where the new length of the tilted stratum, $A''C''$, equals the original length, AC, would $a''B'' = AB$ (Fig. 1). Since the case where this would be true is but one in an infinite number of possible cases, an assumption that the modified and the *original thickness* are the same is equivalent to the assumption that the incompetent strata slipped along their bedding planes freely and without distortion and formed strictly concentric folds.

The Stretching of Plastic Rocks by Vertically Acting Forces.—If, instead of horizontal compression, incompetent strata were deformed by a gradual vertical uplift or marginal depression, they would be stretched and thinned and folds so formed would possess many of the characteristics of ideal similar folds. Under certain conditions simple horizontal thrust might give rise to vertically acting forces for if the amount of crustal shortening were great and affected lower beds their thickening would lift and somewhat stretch those beds near the surface.

It is true that folding under tensional stresses would not result in thickening on the crests and in the troughs, but this minor difference from ideal similar folds would rarely be recognizable in surface exposures, and folds in incompetent rocks due to such agencies as buried faults, compacting of sediments, or isostatic adjustment might well be indistinguishable by the usual local field observations from folds caused by horizontal thrust. Where thrown into similar folds by horizontal thrust, the *vertical thickness* of strata is determined by both the *original thickness* and the ratio of shortening but, where folded by vertically acting forces, the *vertical thickness* of a plastic stratum, $A'B'$ or $A''B''$, would be simply the same as the *original thickness, AB,* irrespective of the dip of the stratum (Fig. 2). The *stratigraphic thickness,* however, would decrease with increasing dip according to the formula, $a'B' = AB \cos \delta'$.

Summary.—Three significant generalizations regarding the deformation of plastic rocks follow from the preceding discussion:

(1) If the axial plane of a fold is vertical, the *vertical thickness* of or the depth to a bed (the distance across a stratum measured parallel to the axial plane) is a constant and depends not at all upon the dip of the stratum but upon its *original thickness* and the ratio of shortening, if any, that resulted from the deformation. If the axial plane were inclined instead of vertical, the vertical distance across the bed would be greater or less than its thickness parallel to the axial plane, depending upon whether the axial plane inclined in the opposite or the same direction as the dip of the bed. (See Fig. 3.) (2) The *stratigraphic thickness* of a bed in a similar fold depends upon the *original thickness*, the ratio of shortening, and the dip. (3) In any plane parallel to the axial plane, all strata in an ideal similar fold will have the same dip.

The writer is well aware that in extreme cases these analyses of ideal similar folds are not applicable to actual conditions. Under the assumptions made, ideally plastic material where folded into dips so steep as to be parallel to the axial plane would have a stratigraphic thickness of zero and overturned folds could not be formed! However, it seems very probable that these deductions approximate the actual behavior of shale beds in most open folds more closely than do those based on the assumption of concentric folds.

DETERMINATION OF THE THICKNESS OF INCOMPETENT BEDS.

Practically all formulae, diagrams, and graphic methods for the determination of the thicknesses of dipping rocks are based on the assumption of parallel or concentric folds,[5] and if incompetent rocks such as shale are deformed into essentially similar folds, thicknesses determined on this assumption may be grossly in error. Simple computations and graphic constructions for determining thickness based on the assumption of similar folds with vertical axial planes very probably give a more accurate thickness

[5] Quite recently an article by E. L. Ickes appeared in the *Amer. Assoc. Petroleum Geologists Bulletin*, vol. 9, pp. 451–463, 1925, on the determination of thickness of strata in similar and other types of folds.

of folded shale beds; they are somewhat simpler than those necessary when concentric folding is assumed; and they give directly the local *vertical thickness* necessary for reducing elevations of outcropping strata to the datum bed used in structure contour maps.

In Fig. 4 is shown a method for determining the *vertical thickness* in similar folds with vertical axial planes where the dip is constant and the outcrops at different elevations. The angle of slope between two stations must first be computed in order to use most of the formulae that have been published but this angle is not determined directly in plane-table mapping. In concentric folds this preliminary determination is not essential, for the horizontal distance and elevation may be so incorporated in the general formula that the slope angle is not required (Fig. 4), and in similar folds with vertical axial planes the difference in elevation of the two outcrops may be directly added or subtracted.

A comparison of the results of the methods based on similar and concentric folding is of interest. From the two equations in Fig. 4, $aB = AA' \tan \delta + A'B$ and $a'B = (AA' \tan \delta + A'B) \cos \delta$, it follows that the relation of the *vertical thickness* determined by the similar folding method to the *stratigraphic thickness* determined by the assumption of concentric folding, $aB/a'B' = 1/\cos \delta$, regardless of the differences in elevation of the two stations. Where the dips are variable the relationship, though more complex, is approximately the same. From the simple relation where the dip is constant, it is determinable that the thicknesses of the beds between stations obtained by the similar folding method are less than 1 per cent. greater than those derived from the concentric folding method where the dips are less than 8°, and less than 5 per cent. greater where the dips are less than 18°. However, the discrepancy between the results of the two methods increases very rapidly with steeper dips and it is precisely in these cases of steeper dips that it is unjustifiable to apply the concentric folding method to incompetent beds.

In the preceding discussion it has been assumed that the dip remains constant; as is well known, in actual cases of either con-

centric or similar folds this rarely happens. Fig. 5 represents the application of the similar folding method to a case where, as is usual, the dips are variable, and is modelled after Hayes'[6]

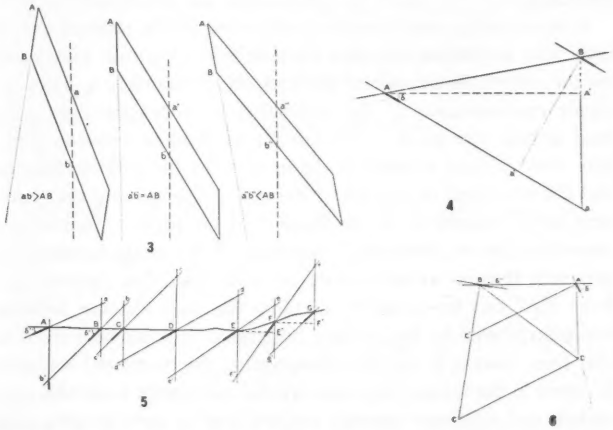


FIG. 3. Effect of inclination of axial plane upon the vertical thickness of a bed.

FIG. 4. Method for determining vertical thickness of beds in a similar fold where the dip is constant and the outcrops are at different elevations. In similar folds: The vertical thickness, $aB = AA' \tan \delta + A'B$. In concentric folds: The stratigraphic thickness, $a'B = (A'a + A'B) \cos \delta = (AA' \tan \delta + A'B) \cos \delta = AA' \sin \delta + A'B \cos \delta$.

FIG. 5. An approximate graphic method for determining the vertical thickness of beds in a similar fold where the dip changes between stations. Vertical thickness of beds

$$A \text{ to } B = \frac{Ab' + aB}{2} = \frac{AB(\tan B + \tan A)}{2};$$

Vertical thickness of beds

$$F \text{ to } G = \frac{Fg' + fG}{2} = \frac{FF'(\tan G + \tan F)}{2} - GF'.$$

FIG. 6. A mathematical method for determining the vertical thickness of beds in a similar fold where the rate of change of dip between stations is uniform.

[6] Hayes, C. W., "Handbook for Field Geologists," pp. 28, 29, 1921.

solution of the thickness of beds in parallel or concentric folds where the dips are variable. The method here proposed is well adapted to a graphic solution and it may also be computed mathematically.

A convenient, quasi-graphic application of the method can be made that somewhat increases the accuracy. Contour lines based on the elevations of several stations are presumably a more accurate representation of the inclination of a stratum than a dip read at only one point. The horizontal distance between structure contour lines is equal to the product of the contour interval and the cotangent of the dip of the bed contoured and the cotangent is the reciprocal of the tangent of an angle. Therefore a determination of the *vertical thickness* of the strata between the two beds that is probably more accurate than that derived only from dips may be found by dividing the map distance between two escarpments by the average horizontal distance between contour lines; that is to say, by extrapolating the contoured surfaces. It seems to the writer that even in ideal concentric folds this convenient and relatively accurate method may be used to advantage if the dips are less than, say, 15°.

This method and its variations, however, must be applied with circumspection for, if the difference between the angles of dip at the two stations is great or if at one of the stations the beds are nearly vertical, the method is very inaccurate. Solutions that are more generally applicable, because more consistent with the premises involved, may be found by a simpler formula if it is assumed that the change of dip between two stations follows the arc of a circle.[7]

From this assumption and two relations brought out above—that in ideal similar folds with vertical axial planes (1) the angle of dip of the strata in any vertical plane and (2) the *vertical thickness* remain constant—we may determine either graphically

[7] See Hewett, D. F., "Measurements of Folded Beds," ECON. GEOL., vol. 15, pp. 367–385, 1920, for a method of determination of thicknesses of strata in concentric folds that is based on this assumption. Due to typographical errors in this article the word *convex* was substituted for *concave* in the statements of the four cases on pages 373, 376, and 377.

or by computation the *vertical thickness* of the beds between two stations.

Given: AB = horizontal distance between stations measured at right angles to the strike (Fig. 6).

δ' = angle of dip of steeper bed,
δ'' = angle of dip of flatter bed,

and that $CB = CD$ = radius of curvature.

Find: AD

$$\angle BCD = \angle BC'A = 180° - (90° + \delta'' + 90° - \delta') = \delta' - \delta'',$$

$$\angle CBD = 90° - \frac{\angle BCD}{2} = 90° - \frac{\delta' - \delta''}{2},$$

$$\angle DBA = \angle CBA - \angle CBD$$

$$= 90° + \delta'' - 90° + \frac{\delta' - \delta''}{2} = \frac{\delta' + \delta''}{2},$$

$$AD = AB \tan \left(\frac{\delta' + \delta''}{2}\right).$$

If station B is at a higher elevation than station A, the difference in their elevations, E, is subtracted; if lower, it is added. The same formula holds for a fold in which the beds are concave upward.

From this is follows that the general formula is

$$t = H \tan \left(\frac{\delta' + \delta''}{2}\right) \pm E,$$

where t = vertical thickness,
H = horizontal distance (at right angles to strike) between stations,
δ' = angle of dip of steeper bed,
δ'' = angle of dip of flatter bed,
E = difference in elevation of two stations; is applied— where the stratigraphically lower bed is higher than the other, + where the stratigraphically lower bed is lower than the other.

This formula, of course, is not precise for, even granting that the folds are similar and have vertical axial planes and that the field data are free from errors, it also assumes that the dip changes uniformly between the two stations. In many places, the variations in the dip are not so simple and where the change is obviously irregular, the field geologist's estimate of the thickness sought may be more accurate than that derived by this or any other formula. The possible inaccuracy or incompleteness of all field data and the uncertainty of our knowledge regarding the behavior of rocks under deformation make it seem to the writer a gross exaggeration to report stratigraphic thickness in figures more definite than, say, one per cent., even when determined from well logs. However, a close approximation to the actual thickness of formations can perhaps usually be found by determining thicknesses of strata as if in concentric [8] and similar folds, and weighting the average of the two figures by the relative amount of the different types of material involved.

THE PRACTICE OF "CORRECTING FOR DIP" IN PREPARING STRUCTURE CONTOUR MAPS.

In making structure contour maps many oil geologists determine the elevation of the datum bed chosen for contouring by applying to the normal stratigraphic interval from the outcropping stratum a factor derived from the observed dip. Several methods of making this correction are in use. Perhaps the most common practice is one that may be called the "Depth to Bed" method.[9] It is based on the assumptions that the *stratigraphic*

[8] For this purpose Hewett's formula for the thickness of beds in concentric folds may be stated without the inconvenient slope angle:

$$t = \frac{H \sin\left(\frac{\delta' + \delta''}{2}\right) \pm E \cos\left(\frac{\delta' + \delta''}{2}\right)}{\cos\left(\frac{\delta' - \delta''}{2}\right)},$$

where all the symbols (except t) and the sign $E \cos\left(\frac{\delta' + \delta''}{2}\right)$ are the same as in the similar folding formula.

[9] Many articles on the determination of the depth to a bed have used this method. Specific mention of its application to the construction of structure contour maps was made by Lahee, F. H., "Field Geology," pp. 415–416, 1916.

thickness is unaffected by the folding and that the observed dip continues downward as well as at right angles to the bedding planes. In Fig. 7 is shown a set of conditions to which the method is obviously not applicable. The line representing *Bed a* is not intended to be the reconstruction of that bed which would be made in a cross section but to indicate the probable shape of a

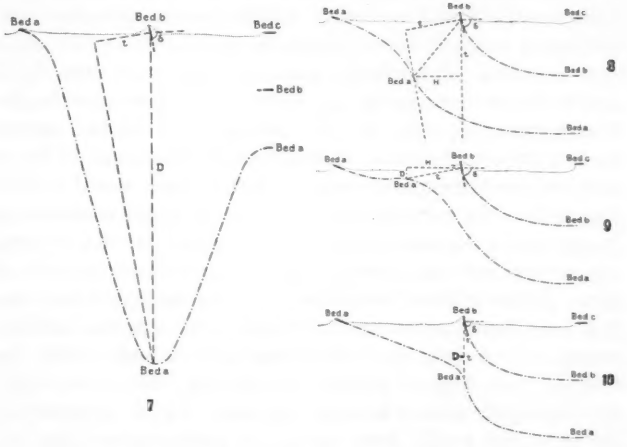


FIG. 7. "Depth to Bed" method of correcting for dip in preparing structure contour maps. Obviously not applicable to the case shown. $D = t/\cos \delta = \sec \delta . t$. Symbols used in Figs. 7, 8, 9, and 10: δ = angle of dip; t = normal stratigraphic interval between *Bed a* and *Bed b*; D = vertical distance to datum bed; H = horiontal distance (at right angles to strike) to point on datum bed.

FIG. 8. "Horizontal Offset" method of correcting for dip in preparing structure contour maps. Makes beds unduly thick where dip is steep. $H = t \tan \delta/2$; $D = t$.

FIG. 9. "Ideal-Concentric Fold" method of correcting for dip in preparing structure contour maps. Laborious, theoretically inapplicable to shale beds, and a possible source of serious errors in the contour map. $H = t \sin \delta$; $D = t \cos \delta$.

FIG. 10. Recommended method (or lack of method) for determining the depth to a bed in similar folds. $D = t$.

surface contoured from the three points on *Bed a* derived from the surface information. The inherent weakness of the method is that it does not permit a variation in the dip from place to place and it is only where the fold is of the concentric type and the dip does not change between stations that the method is applicable.

Another means of applying the correction may be called the "Horizontal Offset" method.[10] Perhaps because somewhat more tedious to apply, it is less frequently used than the "Depth to Bed" method. Precisely the same assumptions are made for it as for the previous method but to the writer it seems to be distinctly preferable. Fig. 8 shows the application of this method to the same set of outcrop observations as those given in Fig. 7 and indicates the probable shape in which *Bed a* would be contoured from the three stations. If the scale of the contour map is such that a horizontal distance of, say, 200 feet is negligible, any correction of less than this amount is obviously a waste of time. If the indicated horizontal offset is greater than 200 feet, it is due either to a greater stratigraphic interval to the bed contoured or to a steep dip,—the very conditions under which the errors in field data are greatest and the applicability of concentric folding to shale beds most doubtful. Fig. 8 demonstrates the error that results from use of the method where dips are variable,—that of making the beds thicker on the limbs than on the crest or in the trough of a fold, a relation which is not true of either ideal similar or concentric folds.

The third process with which the writer is familiar is one that may, for want of a better name, be called the "Ideal Concentric Fold" method.[11] For it the only assumptions made are those inherent in the concept of concentric folds—that the *stratigraphic thickness* does not vary and that the angle of inclination of the

[10] Though many petroleum geologists are probably familiar with this or a similar method, to the writer's knowledge it has not been published.

[11] Like the "Horizontal Offset" method, this method is known to many but so far as the writer is aware it has not been published. In 1923, the U. S. Geological Survey issued to its field employees a mimeographed description of this method, written by K. C. Heald.

strata remains constant at right angles to the dip. Because it necessitates two computations, a scaling of the map, and the subtraction (or addition) of a variable, the method is not widely used, although it is certainly the most consistent, and perhaps the least inaccurate, of the three considered. In Fig. 9, this method is applied to the same outcrop data as those given in Figs. 7 and 8 and the probable shape in which *Bed a* would be contoured from the three points is shown. In this figure the method is not strictly followed for, at the point on *Bed a* derived from the outcrop of *Bed b,* the dip is shown as nearly horizontal instead of the same as at the outcrop of *Bed b* as would be required in strictly concentric folding. This contradiction was made purposely in order to avoid putting an anticline between the outcrops of *Bed a* and *Bed b.* Even with this inconsistency incorporated in the figure, a pronounced structural terrace, not indicated by surface dips, must be drawn between the two stations if the method is applied.

The writer's objections to these methods are that they are unwarranted theoretically and that they may lead to serious errors in the structure contour maps in areas where incompetent beds are steeply folded. The methods may possibly be theoretically applicable and they do not lead to such serious errors in areas of gentler dips, but the extra time required for their application remains a weighty objection to them.

The writer thinks that a return to the old method, or lack of method, of making no such "corrections" whatever is based on premises that are theoretically more probable and that no inconsistencies result when the "corrections" are omitted. That this omission results in a saving of time is certainly not a controversial matter. If shale beds are deformed into similar folds whose axial planes are vertical, and if the beds show no evidence of cleavage or metamorphism (which if present might indicate a high ratio of shortening) it seems justifiable in the preparation of structure contour maps to disregard any effect of the dip on the vertical distance across strata.[12] Fig. 10 shows the application

[12] Assuming, of course, that no direct determinations of the depth of a buried bed have been made by the drill.

of this method to the same set of outcrop observations as those given in Figs. 7, 8, and 9, and the probable shape in which *Bed a* would be contoured from the three stations.

THE RATIO OF SHORTENING.

A Method for Determining the Nature of the Forces that Produced a Similar Fold.—The ordinary field data contain neglected evidence that can be used, supplementary to the regional relations, to indicate whether horizontal thrust was effective in producing a similar fold. Since under vertically acting stresses incompetent beds may be stretched and under horizontally acting forces they may be shortened, the mere length of tilted plastic strata after folding can not be used to determine the amount of horizontal shortening.[13] It is only where the *stratigraphic thickness* is unaffected by the folding—that is, where the strata behave as though competent and form strictly concentric folds—that the length of the strata remains constant during deformation.

However, another method of determining shortening is possible. Although the amount of shortening can not be determined, the ratio of shortening can and this figure is significant evidence as to the origin of a fold. In Fig. 1 the ratio of the *vertical thickness* of the strata after deformation to the *original thickness, A'B'/AB,* or the ratio of the present *stratigraphic thickness* to the product of a trigonometric function of the dip and the *original thickness, a"B"*/cos δ *AB,* equals the ratio of shortening, *AC/A'C'*. A ratio of any given value may be the result of deformation that was intense but local, or gentle but more widespread.

To apply the criterion, the *original thickness* of the strata (determined approximately from several measurements nearby where the beds are relatively unaffected by folding) and their thickness on a fold (determined by averaging several measurements) must be known. In general, a large ratio indicates horizontal thrust; a small one vertically acting forces. If the ratio between these

[13] Van Hise, C. R., "Estimates and Causes of Crustal Shortening," *Jour. Geol.*, vol. 6, pp. 12–17, 1898.

two thicknesses is sufficiently larger than the secant of the angle of dip and the probable error of the measurements, it indicates strongly that the fold was produced in large part by horizontal thrust. If the ratio approximates the secant of the dip, the structure may be either a similar fold due to horizontal thrust or a concentric fold, for in strictly concentric folds the apparent ratio of shortening as determined by this method would be the secant of the dip. If the ratio is less than the secant of the dip and greater than one, the structure may be a similar fold due either to horizontal thrust or, if the axial plane inclines in the opposite direction from the dip of the strata, to vertical uplift (See Fig. 3). The probability of an inclined axial plane may be roughly tested in some cases for the amount of inclination necessary to account for the ratio can be computed from the observed thicknesses and dip. A ratio of approximately one does not necessarily mean that the folds are the result of vertical forces for not all folds due to horizontal thrust would show a high ratio of shortening. Also if the axial plane of a fold inclined markedly in the same direction as the dip of the beds, the apparent ratio would be unduly small (Fig. 3). Even a ratio of less than one does not certainly indicate vertically acting forces for thickening of the lower beds may have been great enough to cause marked stretching of the higher ones (thereby causing a decrease upward of the apparent ratio) or the actual ratio of shortening, due either to slight tangential thrust or great width of the area affected, may have been smaller than the probable error in the field measurements.

Other Possible Uses.—In addition to its bearing on the origin of the fold, the possibility of concentric folding, and the inclination of the axial plane, the ratio of shortening may be used to increase somewhat the accuracy of structure contour maps and to determine the probable order of magnitude of the depth of folding. In the method of preparing structure contour maps illustrated in Fig. 10 the assumption was explicitly made that the shortening was negligible and the interval used at the outcrop of *Bed b* to determine the elevation of *Bed a* was the normal or original interval between them. By an approximate determina-

tion of the ratio of shortening on any fold it is possible to test and even to avoid this assumption.

If the amount of crustal shortening and the width and increased elevation of a deformed block is known, the approximate depth of folding can be computed.[14] However, in similar folds the amount

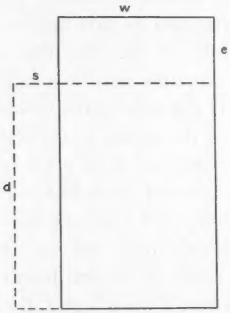


FIG. 11. One method for determining the approximate depth of folding in incompetent rocks. s = shortening during deformation; d = depth of folding; w = width of deformed block; e = elevation due to deformation; $s + w$ = original width of block involved in deformation; $d + e$ = total height of block after deformation;

$$R = \frac{s + w}{w} = \frac{d + e}{d} = \text{ratio of shortening;}$$

$$we = sd.$$

Given: R and e.

Find: d.

$$Rd = d + e$$
$$Rd - d = e$$
$$d(R - 1) = e$$
$$d = \frac{e}{R - 1}.$$

[14] Willis, Bailey, "Physiography and Deformation of the Wenatchee-Chelan District, Cascade Range," U. S. Geol. Survey Prof. Paper 19, pp. 95–97, 1903. The formulæ, which Willis credits to G. F. Becker, use the reciprocal of the ratio of shortening and the average elevation of the block but in their application to the Cascade uplift the amount of the shortening and the width are considered as known quantities. Chamberlin, T. C., and Salisbury, R. D., "Geology," vol. 2, pp. 125–126, 1906. Chamberlain, R. T., "The Appalachian Folds of Central Pennsylvania," *Jour. Geol.*, vol. 18, pp. 228–251, 1910.

of shortening can not be determined from either the length of the tilted strata or from the ratio of shortening (unless the width of the area affected by the deformation is known), but from the ratio of shortening and the average height to which an area has been lifted by deformation a value for the depth of folding can be computed. Fig. 11 illustrates this method. It should be noted that in the diagram as drawn the apparent shortening as determined from the length of the folded strata, $e + w + e - w$, is several times as great as the actual shortening, s, and that the usual formula therefore could not be applied to this case.

This method is based on the assumptions that all shortening was in the plane of the diagram, that it was constant in amount to a certain depth at which it stopped abruptly, that the volume remained constant during the deformation, and that the stress was relieved solely by upfolding. The assumption of uniform shortening to a certain depth may appear unwarranted and it may seem more probable that shortening dies out uniformly downward. With this alternative assumption, there must be substituted for the parallelogram bounded by the sides s and d (Fig. 11) a right triangle whose base equals s and whose altitude equals $2d$ and the formula would be

$$d = \frac{2e}{(R - 1)}$$

In both formulae, the depth of folding, d, increases with increase in elevation, e, and with decrease in ratio of shortening, R. Other assumptions might be made and different values for d determined from them but these two can perhaps be used to indicate the probable order of magnitude of the depth of folding.

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GENESIS OF THE MAGNETITE DEPOSITS NEAR THE WEST COAST OF VANCOUVER ISLAND.[1]

W. L. UGLOW.

CLOSE to the west coast of Vancouver Island, British Columbia, are a number of magnetite deposits of contact metamorphic origin, that have been many times mentioned in geological literature, but have not yet received adequate geological interpretation. These deposits are the subject of a memoir [2] by the writer, now in course of preparation, but therein they are treated from an economic rather than from a scientific point of view. It is the purpose of this article, therefore, to present a brief survey of some of the characteristics of these magnetite occurrences that have a bearing on their genesis.

Vancouver island lies along the west coast of British Columbia, and extends with the northwestern cordilleran trend for a total length of about 250 miles, and an average width of 50 miles. It is deeply indented along its western margin by a number of fjords, locally called sounds, that penetrate, with their finger-like ramifications, from three to thirty-five miles into the heart of the island. It is close to these sounds or natural harbors, that most of the magnetite deposits occur. Along its Pacific slopes, the island is densely timbered and clothed with an almost impenetrable growth of underbrush, and continuous large-scale rock exposures are rare, except along the shores. The backbone of the island is a mountain system whose peaks rise to 7,000 feet above sea level, well above timber line for this latitude.

[1] Published by permission of the Director, Geological Survey, Canada. Presented before Society of Economic Geologists, New York, May, 1925.

[2] To be published by the Geological Survey, Canada.

GEOLOGY.

Dawson,[3] Webster,[4] Kimball,[5] Haycock,[6] Clapp,[7] Dolmage,[8] and others have at various times since 1886 geologically surveyed parts of the Pacific coast of the island, so that the fundamental facts are fairly well established. The following sequence epitomizes the stratigraphy and lithology:

TERTIARY { Lower Miocene?: Conglomerate, sandstone
Lower Oligocene: Sooke gabbro
Upper Eocene: Metchosin basalt, tuff

Unconformity.

Lower Cretaceous: Sandstone, shale, conglomerate, coal

Unconformity.

Upper Jurassic?: Coast range batholith { Saanich granodiorite
Beale diorite

Lower Jurassic or *Upper Triassic* } Vancouver group { Vancouver volcanics and Sutton limestone
Nitinat limestone

Palaeozoic: Carboniferous { Malahat volcanics
Leech river slates

Magnetite occurs at a large number of places in deposits of varied shape and size, but in all cases in areal relationship to the Vancouver series and the intruding diorite and grandiorite. The occurrences are all of contact metamorphic origin and may be classified according to the following system:

(*a*) Magnetite deposits in limestone.
(*b*) Magnetite deposits in andesite and andesitic tuff.
(*c*) Copper-magnetite deposits in limestone, andesite or andesitic tuff and diorite.

[3] G. M. Dawson, Geol. Surv. Can., Ann. Rept., vol. II., 1886, part *B*.
[4] A. Webster, Geol. Surv. Can., Summ. Rept., 1902, pp. 55–76 *A*.
[5] J. P. Kimball, Amer. Geol., vol. XX., 1897, pp. 13–27.
[6] E. Haycock, Geol. Surv. Can., Summ. Rept., 1902, pp. 76–92 *A*.
[7] E. H. Clapp, Geol. Surv. Can., Memoir 13, 1912.
[8] V. Dolmage, Geol. Surv. Can., Summ. Repts., 1918, part *B*; 1919, part *B*; 1920, part *A*.

MAGNETITE DEPOSITS IN LIMESTONE.

The larger and purer deposits occur in crystalline limestone of the Nitinat or Sutton formation near its contact with an intrusive hornblende diorite (called by Clapp the Beale diorite). In some cases, as at the Conqueror deposit, Port San Juan, the contact is nearly vertical and the contact metamorphism small in lateral extent; while in other cases as at the Sarita river deposit, Barkley sound, the contact is an undulating, gently sloping one with the underlying diorite only slightly unroofed, and here the contact metamorphic effects are sporadic but quite extensive. Some of these deposits are mineralizations of limestone inclusions within the diorite, others are mineralizations of limestone roof pendants, while others still are mineralizations of the marginal phase of large bodies of limestone along one side of the diorite intrusion. Massive, non-banded structure is generally characteristic of this magnetite, at least to the naked eye, and many dike-like and sill-like apophyses of magnetite may be observed in the limestone.

MAGNETITE DEPOSITS IN ANDESITE AND ANDESITIC TUFF.

Here again the magnetite occurs as a series of contact deposits, produced by the influence of the Beale hornblende diorite on the invaded porphyrites and tuffs of the Vancouver series. The same variety of occurrences is found in this group as in the group just described, although the distinctions regarding the nature of the contact are not quite so clear. There are steeply dipping (Crown Prince, Barkley sound) and gently dipping contacts (Darby and Joan, Barkley sound), and others in which the contact undulates about a shallow dip (Copper island, Barkley sound). In nearly all cases, these deposits lack the massiveness of the limestone deposits, and are, instead, characterized by a striking banded structure which is a preservation of the stratification in the volcanic tuffs and porphyrites.

COPPER-MAGNETITE DEPOSITS IN LIMESTONE, ANDESITE, TUFF AND DIORITE.

Many copper deposits of contact metamorphic origin are known along this coast, and with most of them magnetite is a common gangue mineral. In some cases (Old Sport mine, Quatsino Sound) abundant magnetite occurs in solid ledges intimately mixed with copper sulphides (chiefly chalcopyrite); in other cases (Tidewater mine, Sydney Inlet) magnetite is disseminated through the deposit and is found in small bunches and stringer-like masses near its base; while in still others, only small quantities of magnetite occur, the chief iron minerals being pyrrhotite and pyrite. At the Tidewater mine the host rock is andesitic tuff; at the Old Sport Mine, it is limestone; while at the June deposit (Quatsino sound) the host is partly tuff and partly Beale diorite. Near some of these deposits the only visible intrusive is diorite; at others both diorite and granodiorite occur; while in other cases still, granodiorite or granite alone is exposed. The evidence afforded by the study of the areal geology points to the ubiquitous nature of the diorite along the coast (even where not exposed) and to the less general or more sporadic occurrence of the granodiorite; and the conclusion is reached that the magnetite mineralization owes its origin to the influence of the diorite, while the chalcopyrite and bornite are to be attributed to the later intrusion of more acidic type—granodiorite or granite.

PARAGENESIS OF THE MINERALS OF THE CONTACT ZONES.

The minerals of the contact zones are, in decreasing order of abundance, garnet (andradite and grossularite), epidote, quartz, calcite, tremolite, actinolite, diopside, zoisite, and the unaltered and altered constituents of the country rocks. There are endomorphic as well as exomorphic changes, and in many cases the results of these two alterations approach one another so closely in mineralogical composition that it is difficult, if not impossible, to locate accurately the position of the intrusive contact.

Where silicate rocks, such as andesite and andesitic tuff are the

host rocks, the magnetite and gangue minerals occur in a mosaic, suggestive of contemporaneous deposition, but where the magnetite occurs in limestone, the evidence is clear that the garnet-epidote zone was of earlier formation and is cut by veinlets and dike-like bodies of magnetite. Many polished surfaces showed magnetite to have replaced garnet.

In all cases, the copper minerals were seen to be of a later generation than the magnetite, and of the copper sulphides, chalcopyrite followed bornite; so that there is proved to be a general sequence of formation of the minerals of the contact zones as follows: (1) silicates, (2) magnetite, (3) bornite, (4) chalcopyrite.

GENESIS.

The restriction of these magnetite deposits to positions close to the intrusive contacts of diorite and granodiorite with limestone and tuff; and their constant association with garnet, epidote, diopside, tremolite or vesuvianite leave little room for doubt as to their ultimate source and the general conditions of their formation. They belong without question to the class of contact metamorphic deposits, for the formation of which important contributions of metalliferous and non-metalliferous constituents are made from the intrusive magma. It is the method of introduction and deposition of the iron salts that formed the magnetite that becomes the essence of this paper.

Clapp's Hypothesis of Magnetite Magma.—During 1909, C. H. Clapp[9] made the first[10] important geological examination of

[9] C. H. Clapp, "Southern Vancouver Island," Geol. Surv. Can., Memoir 13, 1912. C. H. Clapp, Geol. Surv. Cann., Summ. Repts., 1908, 1909, 1910.

[10] An earlier examination of some of these deposits was made by J. P. Kimball in 1897, but in his paper entitled "Secondary Occurrences of Magnetite in Islands of British Columbia, by Replacement of Limestone and by Weathering of Eruptives" (*loc. cit.*), he makes the following statement in reference to genesis: "The occurrences here referred to clearly conform to two separate types of ferriferous deposits which it has seemed important to distinguish as hydro-chemical replacements. . . . One type is morphological replacement of limestone by double decomposition between ferrous salts and calcic carbonate, the former being generated from ferrous silicates; the other type, a partial and not necessarily pseudomorphic, replacement of ferrous silicates in weathered basic rocks, or as more

some of these magnetite deposits—particularly those along Bugaboo creek, and Gordon river, near Port San Juan. Clapp was struck by the rather remarkable occurrence of a bluff of magnetite on the Conqueror claim, of which he wrote the following description: "There the magnetite body, which forms a cliff 30 to 40 feet high over which the creek falls, is in contact with a mass of white crystalline marble. . . . Irregular magnetite veins or apophyses extend from the magnetite body into the pure marble, brecciating it and including blocks of marble much as apophyses of an igneous rock would brecciate and include fragments of an invaded formation." He further elaborates his conception of the origin of the deposit in these terms: "After the limestones had been more or less contact metamorphosed . . . solutions of magnetite, with small amounts of sulphides, penetrated the contact zone, and replaced it in part. The solutions were apparently very concentrated, *virtually magnetite magma,* since they intruded and brecciated the sheared diorite and unaltered marbles in much the same manner as rock magma intrudes and brecciates." [11]

This conception supports the hypothesis so extensively and emphatically advocated by Spurr in recent years for the origin of many metalliferous deposites, and implies the idea of a rapid injection of ore magma with displacement rather than replacement of the country rock as the principal process. After detailed examinations of all of the iron deposits near the west coast of Vancouver island during 1924,[12] the present writer is strongly inclined to disagree with the above hypothesis and to return to the notion of replacement by tenuous solutions.

Replacement by Tenuous Solutions the Dominant Process.—The character of the solutions from which magnetite was deposited is not believed to be magmatic in the sense that they con-

explicitly distinguished, a residual concentration or fixation of iron oxides incidental to development of soluble alkaline carbonates from weathering oxidation or splitting up of ferriferous silicates. In replacements of limestone ferriferous material is generally derived from extraneous though contiguous sources through simple permutations and reactions."

[11] *Loc. cit.*, 1912, pp. 192–193.

[12] Memoir covering the results of this work is forthcoming by the Geological Survey, Canada.

sisted of molten magnetite in a somewhat viscous condition, as postulated by Clapp; and the solutions are not believed to have caused the brecciation of the host rocks. Many examples were studied during the season, and it was observed that, where the host is limestone, the magnetite in many cases takes on a shape resembling dikes, sills or irregular apophyses; but where the host is a siliceous or silicate rock, there is no such tendency. The conclusion is drawn that the dike-like or sill-like bodies of magnetite are not true dikes or sills, but are the results of migration of iron-bearing solutions through previously existing fractured or porous rocks, accompanied by deposition of magnetite within the fractures and the gradual and, in places preferential, replacement of the county rock by magnetite from tenuous solutions of high penetrability. It does not seem at all conclusive to the writer that the iron in the magmatic solutions was even in the form of magnetite; it seems more plausible that it may have been emitted as an iron chloride or fluoride, and precipitated as magnetite by interaction with the rocks.

The hypothesis of replacement by tenuous solutions is based on the following observations:

1. No evidence was found to indicate that magnetite magma actually caused the brecciation of the limestone or tuff, referred to by Clapp, and observed in many other cases. Furthermore, contact zones barren of magnetite or of any solid body of introduced mineral were found to be similarly brecciated; and indeed such brecciation would be expected to be produced at the time of the invasion of the plutonic magma. Why, therefore, is it necessary or advisable to attribute it to a magnetite magma?

2. It is true that, in the examples mentioned, apparently solid magnetite surrounds blocks and fragments of limestone, silicated limestone and diorite, but a similar surrounding of rock fragments with angular outline and sharp boundaries can be accomplished by means of slow replacement of the rock by solutions that worked inwards from open or potential fractures.[13]

[13] J. D. Irving, "Replacement Ore Bodies and Criteria for their Recognition," ECON. GEOL., vol. 6.

A. M. Bateman, "Angular Inclusions and Replacement Deposits," ECON. GEOL., vol. 19, 1924, pp. 504–518.

3. Close observation of the so-called solid magnetite apophyses, with the study of polished surfaces of the same, reveals the fact that they are not by any means solid magnetite, but are impregnated with residual grains of calcite and lime silicates of too small dimensions to be satisfactorily explained as supported xenoliths of country rock. They are unreplaced portions of the contact zone, and in themselves constitute direct evidence of replacement rather than displacement as the principal depositional process.

4. Sections across the contact between these "magnetite apophyses" and the fragments of limestone or silicate rock show that the contact is not a sharp intrusive one, but that it is of the nature of a rapid transition from magnetite with few grains of unreplaced calcite and lime silicates to limestone or silicate rock with a few scattered grains of magnetite. Such a transition is characteristic of replacement deposits.

5. Many of the magnetite bodies, particularly those that occur in limestone, are seen to be completely isolated from the main mass of the magnetite, and are surrounded on all sides by country rock. Even those that are tabular or dike-like in shape are often so isolated. No channel of access for the "viscous magnetite magma" could be found, and the only explanation for their occurrence that the writer can advance is that they were formed by the replacement of the limestone under the direct influence of tenuous solutions of iron salts of high penetrability that travelled (necessarily very slowly) through thin open or potentially open fractures or capillary inter-grain spaces.

6. Sill-like masses of apparently solid magnetite in limestone are common occurrences at the Glengarry-Stormont deposit near Head Bay, Nootka sound, which seem on casual inspection to support the hypothesis of magmatic injection. Close examination of these sheets of magnetite, however, reveals a banding parallel to the walls which is similar to the lamination of the limestone, and believed to be a preservation of it after replacement; while minor amounts of lime silicate minerals and calcite were found disseminated through the magnetite, and point to unreplaced or altered portions of preëxisting limestone.

7. In these sill-like deposits, there are no textural or structural characteristics of igneous sills or dikes, such as fine-grained marginal phases, fluxion structures or the stringing out of xenoliths in zones parallel to the walls.

8. Magnetite deposits in the andesites and andesitic tuffs do not possess the dike- or sill-like shapes, nor are they in other ways suggestive of an origin by magmatic injection. On the other hand, all of their characteristics are strongly indicative of a replacement origin. Strongly laminated structures predominate in the deposits, produced by the interlayering of magnetite zones with zones of quartz and silicate minerals. This structure may be observed to be a preservation, after partial replacement, of the bedding structure of the tuffs or lavas. Even the magnetite zones themselves, which are usually not over one foot thick, are not sill-like but flat lenticular, and have scattered through them grains of the quartz and silicate minerals that are characteristic of the adjacent layers. The magnetite deposit on Copper island, Barkley sound, is a replacement of tuff beds, and well illustrates the characteristics outlined above. Certain horizons in this sedimentary series are characterized by a high magnetite content, and detailed examination showed that these horizons consisted of a series of beds from two to four inches thick, every second one of which consisted largely of quartz, garnet or the original silicates of the tuff. The alternate beds contained the magnetite, not in solid tabular, sill-like masses, but in a number of very flat lenses in parallel orientation, but not touching one another. They are separated by a mosaic of quartz, garnet, epidote and the silicates of the tuff, into which they grade rather rapidly along their boundaries. These flat lenses are themselves impregnated with grains of all of the above minerals.

This is typical replacement structure and could not be formed by injection. The replacement is far from complete, and it is in this respect that these deposits differ most strikingly from those in limestone. In the tuffaceous deposits, at least one half of the host rock is unreplaced and the story of their genesis is more

easily read than in the limestone deposits where almost complete removal of the host rock makes the task of interpretation a difficult one. Since both types of occurrence owe their origins to the same source—contributions from intrusive diorites—the tuffaceous deposits afford a definite clue to the method of deposition of those in limestone, and this is by gradual replacement of the country rock by tenuous solutions of high temperature and considerable penetrability.

It is interesting to note that a more complete replacement of the tuffs by the iron-bearing solutions would have produced a series of tabular or sheet-like magnetite bodies that might wrongly have been interpreted as the result of lit-par-lit injection.

9. In the June deposit, near the southeast arm of Quatsino sound lenses and disseminations of magnetite (associated with chalcopyrite and bornite) are found in a hornblende diorite. The study of polished surfaces of magnetite and thin sections of the diorite from here show that the hornblende has been preferentially replaced by magnetite without any visible channels of connection between the replaced grains. Such a structure bespeaks replacement by tenuous or gaseous solutions.

10. Polished specimens of magnetite, and of mixed magnetite and gangue from many deposits show numbers of minute veinlets of magnetite traversing, in recognized replacement relationships, garnet, epidote, feldspar and other non-metallic minerals of the gangue. These specimens prove that in most of the deposits the formation of the contact silicate zone preceded the introduction of the magnetite which, to a greater or less extent has replaced it.

In addition to the above-mentioned detailed evidence, there are two or three general features that seem to favor the development of these contact metamorphic magnetite deposits by replacement rather than by injection. These are noted below:

1. It is difficult, if not impossible, for a "virtually magnetite magma," which would almost certainly be viscous, to have migrated for any notable distance through very narrow channels in

7. In these sill-like deposits, there are no textural or structural characteristics of igneous sills or dikes, such as fine-grained marginal phases, fluxion structures or the stringing out of xenoliths in zones parallel to the walls.

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In addition to the above-mentioned detailed evidence, there are two or three general features that seem to favor the development of these contact-metamorphic magnetite deposits by replacement rather than by injection. These are noted below:

1. It is difficult, if not impossible, for a "virtually magnetite magma," which would almost certainly be viscous, to have migrated for any notable distance through very narrow channels in

solid rocks, whose temperatures must have been much below that of molten magnetite; and yet the deposits show bodies of magnetite in tiny veinlets quite distant from the igneous source.

2. A dry monomineralic magnetite magma would pass into the solid state at the temperature of solidification of magnetite (1538° C. according to Hodgeman and Large's "Handbook of Chemistry and Physics"). Sills and dikes of molten magnetite could not, therefore, be injected very far into country rock that would in most cases be below such a temperature. On the other hand, it might not require a high content of mineralizers to depress the freezing point of the magnetite melt to the temperature range of formation of contact metamorphic zones. But it is extremely probable that, in order to transform a viscous magnetite melt into a tenuous fluid with the permeating power requisite to explain many features of these deposits, such a great content of mineralizers would be required that the resultant fluid could scarcely be called a magnetite magma.

3. The total absence, as far as is known from present exploration, of magmatic segregation deposits of magnetite in the diorite appeals to the writer as an argument against the origin of the contact deposits by injection of magnetite magma. If the dike-like and sill-like masses of magnetite were of intrusive origin as ultrabasic differentiates of the cooling diorite magma (as implied by Clapp, and as probably would be conceived by Spurr), then why should not such a differentiate settle out by liquidation or crystallization to form magnetite-rich phases of the diorite near its boundaries? As stated above, such have not been found to occur. It is suggested that the almost complete absence of titanium in all of the deposits near the west coast is an argument against the emission from the intrusive of the iron minerals as magmatic magnetite or magnetite magma.

The writer is inclined to go one step farther and ask whether it is not generally true that magmatic segregation deposits of magnetite are somewhat titaniferous and are found in basic rocks.

around the contacts of which contact metamorphic deposits of magnetite are not found; and whether it is not also true that the typical contact metamorphic deposits of magnetite are found around the contacts of igneous rocks of intermediate to higher acidity, with which the segregation type of magnetite deposits does not occur.

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IDENTIFICATION OF COALS.

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In the development of a coal area it is of importance to be able to identify its individual seams. On account of the variation in the more readily noted characteristics of most coal seams, as for example, thickness, purity, and structure of the coal, this is often difficult. The method here suggested for doubtful cases has shown some promise. It is to examine in reflected light under a microscope, etched polished surfaces of coal and to identify if possible the characteristic remains of the imbedded floras. The etching reagent used is selenium oxychloride.

Excellent progress is being made in the identification of coal seams by means of their microscopic plant remains, by various workers and the present paper deals primarily with a different manner of preparing the specimens to be examined.

The study of microscopic plant remains in coal has been accomplished by means of: (1) maceration, (2) microtome sections, (3) thin sections ground on a lap and finished on a hone, (4) polished surfaces treated in various ways, (5) X-ray methods.

In the maceration method, first used by Schulze [1] and later modified by Von Gümbel and others the coal is made to disintegrate by treatment with chemical reagents. The last stages of the reaction are watched under the microscope.

The microtome method has been used by Jeffrey,[2] White, and Theissen [3] and others. The treatment of the coal before cutting requires a long time and the sections produced are of small dimensions, up to four millimeters in diameter.

[1] Schulze, Franz, "Über das vorkommen vohl erhaltenes cellulose in Braunkohle und Steinkohle," *Ber. K. Preuss. Akad. Wiss. Berlin,* 1855.

[2] Jeffrey, E. C., "Anatomy of Woody Plants."

[3] White, David, and Theissen, R. R., "The Origin of Coal," U. S. Bur. Mines Bull. 38, 1913.

The grinding of thin sections is rather an uncertain operation because small grains of hard material as for example pyrite or quartz may become loosened and ruin the thin section. The method has been used successfully by a number of investigators and produces excellent results. Thiessen [4] says that from eight to twelve sections may be made by a man in a day.

The polished surface method consists of cutting and then grinding and polishing one or more surfaces. The simplest means of finishing a coal section is to polish it upon billiard cloth or broadcloth on a lap. The softer ingredients of the coal are abraded leaving the harder materials standing up in slight relief. This result is accomplished by Iwasaki [5] by means of the wet leaves ordinarily used in Japan for scowering and polishing. The customary preparation is to etch slightly the highly polished plane surface of the coal by means of a blowpipe flame [6] or by some chemical reagent such as chromic acid [7] or as is suggested below by selenium oxychloride. Relief of those parts which are more resistent to oxidation is caused by etching either of the flame or by one of the above mentioned chemicals, and the resulting surface is examined under the microscope in reflected light.

Until recently the X-ray investigation of coal blocks has shown but the impurities.[8] Kemp [9] now has impregnated the coal mass with a dense material or solution. If necessary some of the original ingredients of the coal are first removed by means of an appropriate solvent. The distribution of the added substance is

[4] Thiessen, R. R., "Under the Microscope Coal has already Lost Much of its Formed Mystery," *Coal Age*, vol. 18, No. 24, Dec. 9, 1920.

[5] Iwasaki, Chôzô, Tohoku Imp. Univ. Technology Rep., vol. I., 1920.

[6] Turner, Homer G., and Randall, H. R., *Jol. Geol.*, vol. 31, 1923, pp. 306–313.

[7] Seyler, C. A., "Fuel in Science and Practice," 4, 56–66, 1925.

[8] "A Fundamental Study of Japanese Coal," by Chôzô Iwasaki, Tohoku Imperial University Technology Rep., vol. 1, 1920, vol. 2, No. 4, 1922, pp. 235–275.

[9] Kemp, C. Norman, "X-Ray Analysis of Coal," *Colliery Guardian*, vol. 127, No. 3296, Feb. 29, 1924, pp. 539–541. "X-Ray Analysis of Coal and a New X-Ray Examining Unit," *Jol. Soc. Chem. Ind.*, July 18, 1924, pp. 234T–236T. "X-Ray Analysis of Coal with Scientific and Technical Applications," *Inst. Min. Engineers, Trans.*, vol. 67, Pt. 1, Mar., 1924, pp. 59–81.

then determined by X-ray examination. This method shows the woody structure and foreign material within the coal.

The investigation of coal by the use of selenium oxychloride was suggested to the author by Dr. Victor Lenher of the Department of Chemistry of the University of Wisconsin. His kindly suggestions, those of Drs. A. N. Winchell and W. H. Twenhofel and other members of the staff of the Department of Geology of the same institution and the suggestions of Dr. David White, the writer gratefully wishes to acknowledge.

DIAGNOSTIC CHARACTERISTICS

The deposition of coal measures required much time during which the environmental conditions were subject to numerous changes. It has been suggested that even the sandstone strata may differ uniformly one from another to such an extent that refined methods of studying the roof and floor measures may make possible their identification and indirectly that of the adjacent coal seam. The slight success thus far obtained by this procedure, as tested in coal fields, has not yet justified the expenditure of the time necessary for its employment. The time required for the deposition of the seam itself is likely to have been relatively greater than in the case of a similar thickness of the floor or roof. Thus opportunity occurred for variations, not only in the depth of water in the peat-bog, or coastal swamp, where the carbonaceous matter was accumulated, but in the amount of sands or silts added from external sources. These changes in environment encouraged changes in the types of the inhabiting floras. More slowly acting but also having an influence were changes in the prevailing local climate, as for example a change from wet to moderately wet or the reverse. This would affect more particularly conditions of the peat-bog type. Another slow change was plant evolution. The greater the length of time involved the greater the opportunity for these environmental and accompanying plant variations. The net result was a tendency to produce differences between the floras which

have gone to make up the coal of one bed, as compared with those which went into the beds above or below.

These differences, however, especially those connected with the more readily decomposable parts of the plant structures, have been obliterated for the most part as the biochemical process of decomposition progressed at the time of deposition of the plant debris. The chemical compositions of all coalifying masses therefore tend toward uniformity regardless of the composition of the ingredients of which these masses were at first composed.[10] More and more of the characteristic floral structures disappeared until in extreme cases nothing was left but such materials as the resinous or waxy and leathery or horny seed coats, pollen and spore exines and outer substances equally hard to decompose. These were concentrated on account of their greater resistance to decay. White [11] grades plant materials beginning with the less resistant to biochemical decomposition as follows:

1. Protoplasm, sap and meristemic cells, the less protected parenchymatous tissues; fleshy parts of the fruits; proteins, starches, sugars, largely composed of water with a small amount of cellulose matter, mostly carbohydrates.

2. Pitch, thick-walled parenchymatous cells, mesophyll, lacunose tissue, bark, cortex, pericycle, mucus, pectines, etc.; thin-walled elements with watery protoplasms, and less resistant celluloses, carbohydrates.

3. More resistant celluloses, pectose, cutocelluose, gums, oils, seed food in storage, mainly carbohydrates with some nitrogen and a little sulphur.

4. Schlerenchymatous elements, cutinized cell walls, less resistant envelopes of seeds, cone scales, etc., some carbohydrates and some special secretions lower in oxygen and higher in hydrogen.

5. Some epidermis, cuticles, schlerchymatous leaf bases, less resistant seed coats and fats, in part carbohydrates, but as a whole much higher in hydrogen and much lower in oxygen.

6. Resinous or waxy and leathery or horny seed coats; waxy integuments and other waxy matter; most resistant cuticles of special composition; sporangia walls; waxes, pollen and spore exines, resins, chitinous matter and charcoal; largely resinoid waxes or "higher" fatty or chitin-

[10] "Environmental Conditions of Deposition of Coal," David White, *Trans. of the Amer. Inst. of Mining and Metallurgical Engineers*, vol. 71, Feb., 1925.

[11] White, David, "Carbonaceous Sediments." Unpublished manuscript in the possession of Dr. W. H. Twenhofel.

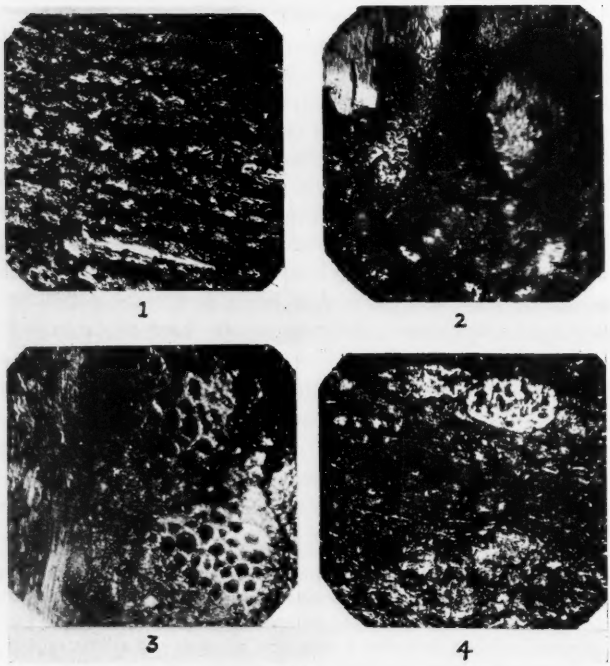

FIG. 1. Micro-photograph of etched polished surface of Manchurian Jet. ×85.

FIG. 2. Micro-photograph of etched polished surface of resin from Manchurian coal. Most attacked part expanded and is out of focus. ×85.

FIG. 3. Micro-photograph of part of section shown in Fig. 5, showing detail. ×85.

FIG. 4. Micro-photograph of etched polished surface of bituminous coal showing spore structure. From Cumberland, B. C. Depth 1,392 ft. ×85.

ous materials, very high in hydrogen and very low in oxygen. Most of these substances are apparently indestructible in the peat-forming medium at any degree of concentration of the humic decomposition products.

Further obliteration of characteristic plant remains may have taken place as a result of later dynamochemical alteration but the presence of some of the materials of group 6 may be depended upon to aid in any microscopic determination of the identity of a coal seam, and spore exines are particularly useful for diagnostic purposes. The more easily perishable plant parts above mentioned may be found in coals in the deposition of which decay of the plant matter was checked at an early stage, and in "coal balls."

CHARACTERISTICS OF SELENIUM OXYCHLORIDE REACTIONS.

The treatment designated to bring into relief the details of such plant anatomy as have been preserved in the coals is one of further differential decomposition by means of the chlorinating oxidizing reagent selenium oxychloride. The following information concerning this acid draws freely upon the articles written by Lenher [12] and Schrenk.[13]

The action of selenium oxychloride is quite selective. Reaction takes place with all the unsaturated hydrocarbons and is explosive with those of the amylene type. The hydrocarbons of the aromatic class such as benzene and toluene form physical mixtures, miscible in all proportions, and may be used as diluents, as well as carbontetrachloride, chloroform and carbon disulphide.

The saturated aliphatic hydrocarbons, at ordinary temperatures are not acted upon or at most are but little affected. Dry cellulose, as for example wood, cotton, or filter paper, is not altered by selenium oxychloride.

The reaction of selenium oxychloride upon coal is complex

[12] Lenher, Victor, *Jour. Am. Chem. Soc.*, vol. XLIII., No. 1, Jan., 1921; and vol. XLIV., No. 8, Aug., 1921.

[13] Schrenk, W. T., Thesis in the Library of the University of Wisconsin, "The Action of Selenium Oxychloride on the Different Coals."

and products of unknown composition are formed. The reagent probably dissolves some of the constituents and reacts chemically with others. It is probable that some oxidation takes place during the extraction. The reaction varies with the amount of volatile matter present in the coal. The bituminous coals have been dissolved to the extent of from 10 to 20 per cent. of their weight. The lower grades are more effected than those at the top of their class. Anthracite containing practically no volatile matter is little acted upon, and thoroughly ignited coke loses no weight by the treatment. Otherwise stated the reaction with anhydrous coaly materials becomes less with each advance through the divisions of White's scale given above, or the action upon the various plant remains is of somewhat the same order of magnitude as that of natural decomposition. The reaction of the acid may be hastened by the application of heat and is entirely destroyed by the addition of water as hydrolysis takes place.

THE PREPARATION OF SAMPLES.

Coal samples may be prepared for treatment with selenium oxychloride as follows:

1. Rough grind on an iron lap using No. 120 carborundum powder with running water.
2. Grind on an iron lap using No. 3 F carborundum powder with running water.
3. Hand grind on a piece of plate glass, using No. 303 optical emery.
4. Grind on a No. 3 F carborundum lap using a gentle stream of water.
5. Dry both the coal sample and the lap till they feel dry to the hand.
6. Polish dry on the 3 F carborundum lap.
7. Hand grind on a hone if a better polish is desired.

By this means samples two inches or more in diameter may be polished. In routine work, where the final hone polishing is not necessary, a specimen may be prepared in ten minutes. This class of work will produce a relatively plane and polished surface which for many of the bituminous coals and for the cannel and anthracite coals is a mirror sufficiently good to give a distinct image of a man 50 yards away.

5

6

7

FIG. 5. Micro-photograph of etched polished surface of bituminous coal from Cumberland, B. C., from a depth of 792 ft. Shows cell structure of woody constituents. × 40.

FIG. 6. Micro-photograph showing another part of specimen illustrated in Fig. 5. Note oval areas that apparently are resin. × 180.

FIG. 7. Micro-photograph of etched polished surface of bituminous coal from Coal River, West Virginia, illustrating spore and cell structure. × 150.

Note that Figures 5, 3, 6 and 4 form a series illustrating the feasibility of distinguishing one seam of coal from another.

TREATMENT OF SAMPLES.

Only anhydrous selenium oxychloride should be used. The sample to be etched should be air dried after the polishing operation. It is lifted by means of glass tongs and placed polished side down in the etching liquid, contained in a stoppered weighing bottle. Either chloroform or carbontetrachloride may be used as a diluent for the reagent, the former permitting the more rapid reaction for the same strength of solution. In the case of anthracite coal the pure reagent should be used. Even with this an hour or more may be necessary to produce sufficient etching. For the subbituminous coals a strong reagent is found to be desirable and a mixture consisting of approximately three parts of selenium oxychloride to one of the diluting liquid is suitable. For ordinary bituminous coals a 50 per cent. solution is used. No special care is taken to insure mixtures of a known strength as the reagent is used in small quantities and is constantly weakening owing to its reaction with the constituents of the coal. In routine work it is sometimes found desirable to have a specimen in each of two weighing bottles at the same time, but greater capacity than this is seldom needed owing to the speed with which the reagent works and the varying nature of the samples treated, which necessitates constant observation of the progress of the etching. When it is desirable to examine the specimen it is removed from the reagent by means of the glass tongs and rinsed for a moment in a quantity of the diluting liquid, then quickly examined under a hand lens. If the surface is covered by an oily-looking liquid, as is often the case, or if for any reason the detail of the etching cannot be observed, the sample is rinsed in hydrochloric acid and then oxidized in a potassium permanganate solution. After having been cleared by this treatment it can be examined without difficulty.

Up to this stage the sample should be kept away from the air as much as possible to prevent the formation of red selenium which is likely to collect on the surface of the specimen and which is difficult to remove. The rather thorough oxidation just de-

scribed tends to remove any of this material which may have formed, and it should be continued until the surface of the sample is once more clear.

If further etching is necessary the sample may be washed with alcohol for a few moments, to hasten drying. It may then be dried in an air jet and immersed again in the etching solution after which the treatment described above should be repeated. When the sample is sufficiently etched it is oxidized as before, washed in running water and allowed to dry.

Severe etchings were not found to be advantageous. The polish upon that part of the coal which is attacked by the acid is destroyed and the underlying material expands and causes some relief. The amount of this relief should be very small for high power magnification with the metallographic microscope.

Coals contain three types of ingredients, namely:

1. Dull coal, more or less granular owing to decomposition having left the more resistant parts.
2. Bright coal which is composed of the woody material, the anthraxylon usually in lenses, that part in which decomposition has not taken place to the same extent as in the dull coal.
3. Mineral charcoal, "mother of coal" or fusain which is the less altered woody material.

The reagent attacks some of the constituents of the dull coal first, the bright coal next and the fusain the last of all. Bituminous coals are readily etched and a sample may usually be treated in from five to ten minutes. Some of the subbituminous coals are etched with much more difficulty. It is supposed that the content of moisture and perhaps of resin may be the cause of this. Resins have been classed by White in group 6, of his scale of decomposition. They are etched slowly by selenium oxychloride. A sample of bituminous coal from the Cretaceous measures of Cumberland, British Columbia upon etching left exposed the circular ends of resinous rods of an average diameter of about 0.8 or 0.9 mm. These rods doubtless filled resin pores in the wood and their presence and characteristics are useful for diagnostic purposes. Anthracites are etched slowly unless the reaction is hastened by heat.

The accompanying illustrations are all from micro-photographs of polished surfaces etched with selenium oxychloride.

The characteristics of the treated sections from which the

illustrations have been selected are sufficiently diagnostic to give an investigator a good idea of the type of coal under consideration and appear to be sufficient to enable him from suitable specimens to distinguish, in most instances, one seam of coal from those lying above or below it.

CONCLUSIONS.

The selenium oxychloride method of treatment of polished surfaces of coal offers a quick and easily controlled means of bringing into relief, delicate detail of plant structure. As a means of treating coal surfaces in order to distinguish one seam from another it offers much promise.

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MAGNETITE–MARTITE–HEMATITE.

JOHN W. GRUNER.

INTRODUCTION.

THE relationship of the iron oxides to one another has been discussed frequently and the problem of the resistance of magnetite to oxidation, especially, has led to controversies.[1]

The writer made some simple experiments in order to find out how magnetites from different localities and of different origin behave toward oxidation in air. Some of the experiments with solutions were made by Mr. D. E. Wachtel.

EXPERIMENTS.

Upon the writer's suggestion he tried to oxidize magnetite by subjecting its powder (200 mesh) to the spray of boiling water and the spray of boiling molar solutions of NaCl, Na_2CO_3, and $MgSO_4$ in flasks. These experiments were continued for four months without noticeable results. Since chemical reaction at 100° C. is probably 500 times as rapid as at 10° C., in four months about as much reaction should have occurred as in 200 years at a temperature of a 10° C. under wet conditions.

The writer then subjected polished specimens of magnetite with some gangue to temperatures of 140° to 150° C. and 190° to 200° C., respectively, in large ovens (as used in bacteriological work). Air could circulate freely. In most cases the specimens

[1] Lindgren, W., and Ross, C. P., "The Iron Deposits of Daiquiri, Cuba," *Trans. Am. Inst. Min. Eng.*, vol. 53, pp. 40–59, 1916, with discussion by L. C. Graton and others.

Roesler, M., "Iron-ore Deposits of the Firmeza District," *Trans. Am. Inst. Min. Eng.*, vol. 56, pp. 77–127, 1917.

Broderick, T. M., "Some of the Relations of Magnetite and Hematite," ECON. GEOL., vol. 14, pp. 353–366, 1919.

Gruner, J. W., "Paragenesis of the Martite Ore Bodies and Magnetites of the Mesabi Range," ECON. GEOL., vol. 17, pp. 1–14, 1922.

were broken in two (parts *A* and *B*, Table I.), and each part was polished. A suitable area in each part was photographed and its position on the stage of the metallograph was recorded. Then the specimens were put in the ovens, *A* specimens at 140°–150° C., *B* specimens at 190°–200° C. They were examined at first every few days for oxidation products. Later examinations were made about once every two weeks, and after two months every four weeks. Since the specimens would show a very slight tarnish after heating for long periods, they were held on the polishing wheel for a second before reëxamination. Though the tarnish was probably due to oxidation, it was so slight and almost invisible that it becomes insignificant in comparison with other changes. It was iridescent in nature and seemed to have no relation to the structure of the magnetite grains. The same areas of the specimens were examined every time with and without oil immersion.

The following table shows the time of the beginning of visible oxidation to hematite at the two different temperatures and the estimated percentage of the visible surface area of magnetite converted to hematite after 330 days. The origin of the deposit from which the specimen comes is given, as well as the amount of hematite it contained before heating. As far as possible specimens without any hematite were selected, but, as the table shows, they are rare. This fact has been mentioned before.[2] However, practically none of the hematite in the specimens before heating occurred as laths along octahedral planes of the magnetite. It always was present as irregular grains, patches, or specks in the magnetite.

It will surprise one, therefore, that *none* of the hematite formed in the process of heating ever had the shape of such irregular grains, patches, or specks. It *always* formed along the octahedral planes (111) of magnetite. It would begin as minute lines barely visible with highest magnification, which would grow longer and a little wider until they finally formed a pattern of triangles so

[2] Broderick, T. M., *op. cit.*
Gruner, J. W., *op. cit.*
Gilbert, G., "Some Magnetite-Hematite Relations," ECON. GEOL., vol. 20, pp. 592–593, 1925.

TABLE I.

OXIDATION OF MAGNETITE TO HEMATITE BY HEATING IN AIR FOR 330 DAYS.

Temperature 140°–150° C.

Number.	Locality.	Type of Deposit.	Size of Grains. Amount of Hematite.	Oxidation Begins After.	Per Cent of Total Oxidation.
B-8	Daiquiri, Cuba	Contact met.?	Medium grained Specks of Hem.	40 days	15–20
M695A	Mesabi Range Biwabik form. Drill core	Sedimentary	Very fine grained A little Hem.	60 "	20+
M1030A	Vermilion Range Soudan form.	Sedimentary	Very fine grained Specks of Hem.	75 "	20+
M999A	Vermilion Range Soudan form.	Sedimentary	Very fine grained A little Hem.	95 "	20+
M2612A	Cuyuna Range Deerwood form. Drill core	Magn. schist Sedimentary	Fine grained Some Hem.	225 "	10–15
M1011A	Vermilion Range Soudan form.	Contact action on sedimentary	Medium grained	275 "	5–10
M2800	Vermilion Range Soudan form.	Contact action on sedimentary	Medium grained	275 "	10–15
101A	Mineville, N. Y.	Magmatic seg.?	Medium grained Specks of Hem.	275 "	5–10
M1251A	Cuyuna Range Gabbro. Drill core	Magmatic seg.	Medium to fine Ilmenite and a little Hem.	300 "	10–20
M282A	East Mesabi R. Drill core	Sedimentary and contact action	Fine to medium	300 "	5–10
398A	Fairview, N. M.	Contact met.	Medium grained Much Hem.	—	None
102A	Magnet Cove, Ark.	Magmatic seg.	Coarse grained A little Hem.	—	None
		Temperature 190°–200° C.			
M709	Mesabi Range Biwabik form. Drill core	Sedimentary	Very fine grained	3 days	Not heated longer
M695B	See above	See above	See above	3 "	40+
M1030B				8 "	25+
M999B				12 "	25+
M2612B				40 "	25+
M1011B				75 "	20+
M1251B				75 "	20+
101B				95 "	10+
M282B				95 "	20+
398B				300 "	5–10
102B				—	None

well known in the replacement of octahedral solids (Pl. I., *A*). In the most oxidized specimens the lines of hematite became quite thick and probably would have covered the whole surface if left in the oven. There was no particular favoring of any areas around the grains of original hematite, nor any growth whatsoever of the original grains. The immediate contacts between the magnetite and gangue minerals showed no oxidation either, unless the contact was an octahedral plane of magnetite. In that case oxidation might have produced a fine line of hematite at the contact.

Pyrite in specimen M-1251 *A*, M-1251 *B*, M-2800, and B-8 remained perfectly fresh. Ilmenite, which was very abundant in M-1251 *A* and M-1251 *B*, remained unaltered, of course, and could easily be told from magnetite (ordinarily not possible unless etched with hot concentrated HCl) as soon as magnetite began to oxidize (Pl. I., *B*). One of the most instructive phases of the experiments is that polished surfaces of the magnetite grains did not seem to oxidize (excepting the tarnish mentioned) unless these surfaces were exactly parallel to the octahedral planes. An experiment (not mentioned in the table) to prove this point was made with an almost perfect octahedron of magnetite from Port Henry, New York. One plane of the octahedron was polished and then subjected to the same treatment as the specimens in the over at 190° to 200° C. It showed no network of oxidation, although there seemed to be a slight change of color over the whole surface. Its electrical conductivity remained that of magnetite. The evidence, therefore, is not conclusive either way.

The table shows a surprising disagreement in the rate of oxidation and the time when oxidation became first visible under the microscope. The difference in the rate of oxidation becomes enormous when the depth to which oxidation penetrates in the specimen is investigated. Four specimens were examined for depth by breaking the pieces in the middle and polishing planes at right angles to the originally polished planes. In three specimens it was found that the amount of oxidation in the grains diminishes very rapidly with depth without, however, showing

any relation to gangue, cracks, or grain boundaries. In specimens M-1251 *B* and 101 *B* oxidation did not penetrate more than 1/25 mm., if that much. Specimen M-999 *B* showed oxidation to a depth of more than 2 mm., though slight in amount, while specimen B-8 again showed the largest quantity of oxidation regardless of depth. (The specimen had a least diameter of 12 mm.)

These measurements show that oxidation, instead of being only 5 to 10 times more rapid, may be hundreds of times faster in some magnetites than in others.

With the exception of B-8, which is probably a contact metamorphic magnetite,[3] it appears that the magnetites of sedimentary origin are more susceptible to oxidation than those of magmatic segregation and contact metamorphism. Of the sedimentary ones, those of the Biwabik formation of the Mesabi range (East Mesabi excluded) are changed most rapidly. Then those of the Soudan formation of the Vermilion range follow, while the magnetite schists of the Cuyuna range (Deerwood formation) are last. One would expect the magnetite of the Soudan to be as resistant as the Cuyuna schist, but the Soudan specimens are a little finer in grain and the Cuyuna magnetite appears to be close to a gabbro intrusion.

The East Mesabi range specimens oxidize much more slowly for the same reasons. The Soudan specimens M-1011 and M-2800 are coarser grained than the others from the same range, and intimately connected with intrusive porphyries in the iron formation, which would explain their relatively slow rate of oxidation.

The size of the grain seems to have great influence on the beginning and rate of oxidation. The rule is the smaller the grain the earlier and faster the oxidation. Daiquiri magnetite is the only one which oxidizes much faster than would be expected. It shows an anomalous behavior in this respect. The reason for the earlier oxidation of the smaller grains is not clear at first,

[3] Lindgren, W., and Ross, C. P., *op. cit.*

considering that the surfaces are polished and the large grains can be attacked as easily as the small ones.

Metallographers speak of diffusion of solids in solids in connection with growth of relatively large grains at the expense of small ones. They have found that diffusion from one grain to another is more rapid, the smaller the size of the grains.[4] Similar reasoning may be applied in the case under consideration. Oxidation is diffusion of oxygen atoms in the magnetite space lattice, as will be shown later. Small grains have a larger surface energy than large ones. (Compare with crystals in liquids.) They have the tendency to decrease this energy, if possible, by growth or other means of adjustment. In metals this growth of grains begins when the temperature is raised to a point where the space lattice permits relatively more rapid diffusion of atoms due to its expansion. In minerals the same principle should apply to grain growth. In magnetite a temperature of 200° C. under ordinary pressure and oxidizing conditions would hardly be sufficient to cause grain growth. The surface energy of the grains, however, can be decreased by oxidation to hematite, for this mineral probably has a smaller surface energy than magnetite under the conditions stated, as will become apparent after the discussion of their structures.

STRUCTURE OF MAGNETITE.

The study of crystal structures by X-rays is of wide application in geology.[5] How it may be used in the interpretation of the experiments on magnetite will be shown below. Magnetite crystallizes in the holohedral class of the isometric system, as shown by Bragg and Nishikawa.[6] For the present discussion the writer constructed a model which represented a unit cube or cell of

[4] Jeffries, Zay, "The Trend in the Science of Metals," *Trans. Am. Inst. Min. Eng.*, No. 1337-N., p. 8, 1924.

[5] Kerr, Paul, "The Determination of Opaque Ore Minerals by X-ray Diffraction Patterns," ECON. GEOL., vol. 19, pp. 1–34, 1924.

[6] Bragg, W. H., "The Structure of the Spinel Group of Crystals," *Phil. Mag.*, vol. 30, pp. 305–315, 1915.

Nishikawa, S., "The Structure of Some Crystals of the Spinel Group," *Proc. Math. Phys. Soc. Tokyo*, vol. 8, p. 199, 1915.

magnetite containing eight molecules of Fe_3O_4 (Fig. 1). In counting the atoms it must be remembered that each one at a corner of the cube belongs to eight cubes meeting in this corner and

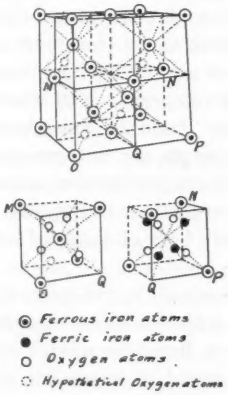

FIG. 1. Unit cube of magnetite. For simplicity's sake only the ferrous atoms are shown in the large cube. Each small cube fits into the octant with corresponding letters and into three more alternate octants in the large cube.

can be counted as only one eighth atom, while those in the faces of the cube can be counted only as half atoms, the other halves belonging to adjacent cubes. The length of the unit cube of magnetite is 8.30 Å. (angstrom units). In other words, a row of 12 million cubes would be about one centimeter long. The circles in the figures represent the centers (really the centers of gravity) of the atoms. They are really ions having lost or gained an electron depending upon their charge, but for the present discussion the word atom will be more convenient.

It can be shown that the space lattice has planes and glide planes of symmetry, axes and screw axes of symmetry, and also centers of symmetry which place it in the highest class of the isometric system. Such a discussion, however, is beyond the scope of

this paper. It will be noticed that each ferrous atom is surrounded by four equidistant O atoms (spacing 1.79 + Å), and each ferric atom by six equidistant O atoms. The spacing is a little larger between the O and ferric atoms (2.075 Å). The O to O distance is uniform throughout and equal to 2.92 + Å.

In what manner these atoms (ions) are connected by bonds is not quite clear. There are two kinds of bonds possible in crystals, depending upon their composition and other factors, polar bonds and "electron sharing" bonds.[7] Polar bonds exist between elements of opposite charges, as, for example, Fe and O, while "electron sharing" bonds are between atoms of the same charge (for example in native elements) in crystals though this term has been used by Langmuir[8] for all kinds of compounds in his octet theory.

Polar bonds in crystals are not to be confused with those bonds we usually associate with molecules, regardless of whether these molecules are gaseous, liquid or solid. As can be seen in the structure models, it would be impossible to separate the crystal into distinct molecules, (this statement does not include organic crystals), since each atom (ion) is held by *all* its neighbors by electrostatic forces. Nor is each bond between an iron and oxygen atom in magnetite, for example, comparable to the valency bond of the old conception. If this old idea were possible, each ferric atom with six equidistant neighbors would have a valence of six in the case under consideration. Each atom (or ion) may be surrounded by any number of oppositely charged atoms as long as the electrostatic equilibrium is maintained in the lattice, for as the Braggs[9] express it, "these forces (which maintain the equilibrium) correspond to a valency bond which can be subdivided to any extent."

[7] Bragg, W. H., and Bragg, W. L., "X-rays and Crystal Structure," fifth edition, 1925, p. 155.

Lewis, G. N., "The Atom and the Molecule," *Jour. Am. Chem. Soc.*, vol. 38, pp. 762–785, 1916.

Langmuir, Irving, "The Arrangement of Electrons in Atoms and Molecules," *Jour. Am. Chem. Soc.*, vol. 41, pp. 868–934, 1919.

[8] Langmuir, *op. cit.*

[9] Bragg, W. H., and Bragg, W. L., *op. cit.*, p. 155.

However, phenomena like magnetism and electrical conductivity are not explained by such bonds. Conductivity is supposed to be due to free electrons in solids,[10] but so far conclusive proof is lacking. Magnetite is an excellent conductor, while hematite is a relatively very poor conductor.[11] On account of the ferrous iron, magnetite has fewer O atoms than hematite. The ferrous atoms are held by fewer polar bonds than the ferric, though all seem to be in equilibrium. Where, then, is the difference between the ferrous and the ferric atom or ion? According to Thomson,[12] it is in the arrangement of the electrons, the ferrous iron having two, the ferric one three electrons in its outer layer. The ferrous one, however, would have one electron in one of the inner layers in excess of the required number. The present writer thinks that this extra electron might account for the conductivity. It certainly does for the instability of the ferrous ion, for "at the first chance" (oxidizing conditions at higher temperatures) this electron will move into the outer layer and will be captured by an oxygen atom which has a capacity for two extra electrons. Such electrons and electrostatic bonds offer difficulties in the explaining of crystal faces, for it seems obvious that unsatisfied forces and "loose" bonds must be on these faces, if the space lattice is terminated abruptly along some planes. This free surface energy tries to approach a minimum value in crystal growth (analogous to surface tension in liquids), which results in faces with the simplest indices.[13] These considerations must also apply to cleaveage of crystals. It should be parallel to planes which have the least free surface energy; or, stated in another way, it should cut the smallest possible number of bonds. Wulff[14] has

[10] Königsberger, J., and Schilling, K., "Ueber Elektrizitätsleitung in festen Elementen und Verbindungen," *Ann. d. Phys.*, vol. 32, pp. 205–209, 1910.

[11] Bäckström, H., "Elektrisches und thermisches Leitungsvermögen des Eisenglanz," *Z. f. Kryst.*, vol. 17, p. 424, 1890.

[12] Thomson, J. J., "The Electron in Chemistry," Philadelphia, 1923, p. 98.

[13] Yamada, M., "Ueber die Oberflächenenergie der Kristalle und die Kristallformen," *Phys. Zeitsch.*, vol. 24, pp. 364–372, 1923.

[14] Wulff, Georg, "Ueber die Natur der Spaltbarkeit der Kristalle, *Phys. Zeitsch.*, vol. 21, p. 718, 1920.

See also Beckenkamp, J., "Atomanordnung und Spaltbarkeit," *Z. f. Krist.*, Festband 1923, pp. 7–39.

shown on a number of minerals that this is true. Let us apply these principles to magnetite. The planes parallel to the cube faces (100) have the spacing and arrangement shown in Fig. 2a.

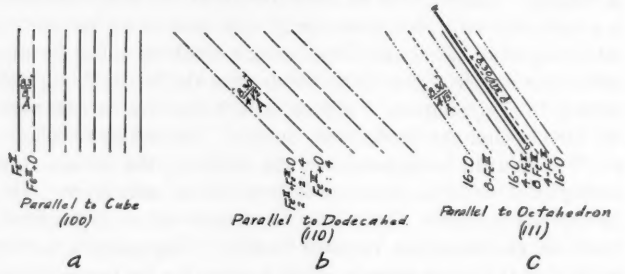

FIG. 2. Diagram showing arrangement and spacing of atomic planes in magnetite. Line *a-b* is along plane of weakness.

Planes of ferrous atoms alternate with planes containing ferric and O atoms. The intensity of bonds between these planes must be great. The planes (110) connecting diagonally opposite edges of the cube (rhombic dodecahedron) are spaced and alternate as shown in Fig. 2b. Bonds between these planes are not so strong because iron and oxygen are found in both sets though in a different ratio. (See numbers under elements in Fig. 2 *b*). These dodecahedral planes form faces, though generally small, in many crystals of magnetite. The planes parallel to the octahedron (111) are spaced and grouped as is shown in Fig. 2 *c*. Many strong bonds hold the ferric atom planes to the O planes (referring to those spaced $\frac{1}{4} \times (8.30/\sqrt{3})$ Å.). However, between two oxygen planes there are three planes with nothing but iron atoms (spacing $\frac{1}{8} \times (8.30/\sqrt{3})$ Å.). The weakest bonds, though closest spacing, should be between two of the iron atom planes (see line *a-b* in Fig. 2*c*). Magnetite, therefore, should develop the octahedral faces most prominently, and if possessing cleavage this should be parallel to these planes. These considerations are born out by the facts.

STRUCTURE OF HEMATITE.

Exact measurements of the structure of hematite have been made lately by Paulding and Hendricks.[15] Fig. 3 represents a

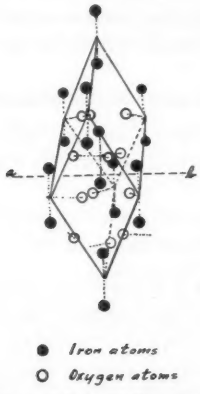

FIG. 3. Unit rhombohedron of hematite. The Fe atoms are spaced dumb-bell fashion on the corners and in the center of the rhombohedron. Three O atoms may be imagined grouped around the center of gravity of each dumb-bell.

drawing of hematite from a model made by the writer. The rhombohedral unit cell has been outlined. It contains 2 molecules of Fe_2O_3. The edge of the rhombohedron is 5.420 Å. and the angle $\alpha = 55° 17'$. Each ferric atom is surrounded by six O atoms, which are *not* at the corners of a regular octahedron as in magnetite. Three are a few per cent nearer the metal atom (1.985 Å.) than the other three (2.060 Å.). Each O atom is surrounded by four metal atoms, two of which are nearer than the other two. This arrangement gives us two values for the O to O distances, the smaller of which is 2.545 Å., which is considerably less than in magnetite. (2.92 Å.). Hematite has a tabular habit parallel to the base (111), as can be seen especially

[15] Paulding, L., and Hendricks, S. B., "Crystal Structures of Hematite and Corundum," *Jour. Am. Chem. Soc.*, vol. 47, pp. 781–789, 1925.

in micaceous hematite. This habit can be explained by relatively weak bonds between two planes of Fe atoms (plane along line *a-b* in Fig. 3). Though these two planes of iron atoms are close together they have the same number of atoms per unit area (reticular density), and each plane can be satisfied by the adjoining O plane, an ideal condition for planes of small free surface energy.

VOLUME AND MASS RELATIONS OF MAGNETITE AND HEMATITE.

Knowing the volumes of the unit cells[16] of the minerals the theoretical densities or specific gravities may be calculated. They are:

$$\text{density of magnetite} = 5.345 +$$
$$\text{density of hematite} = 5.257 +$$

For geological work it is important to know how much increase in volume to expect when magnetite is oxidized to hematite by addition of oxygen. Since volume and contents of the unit cell of each are known, the volume increase may be figured in the following manner. Two molecules of Fe_3O_4 occupy

$$\frac{571.80 \times 10^{-24}\text{ cc.}}{4} = 142.947 \times 10^{-24}\text{ cc.}$$

Hematite with the same quantity of Fe, *i.e.*, three molecules of Fe_2O_3, occupies 150.36×10^{-24} cc., or approximately 7.41×10^{-24} cc. more space. We, therefore, have the proportion $142.947 : 150.36 = 100 : x$. The increase of volume in oxidation is, therefore, 5.2 per cent. This is more than twice as much as one would expect from the density relations.[17]

THE MECHANISM OF OXIDATION.

In order that magnetite become hematite, O atoms must be introduced in the lattice. This O seeks the ferrous atoms which now become ferric. The only reasonable positions for the four

[16] Vol. of unit cell of magnetite $= 571.80 \times 10^{-24}$ cc.
Vol. of unit cell of hematite $= 100.24 \times 10^{-24}$ cc.

[17] Van Hise, C. R., "A Treatise of Metamorphism," U. S. Geol. Survey Mon. 47, p. 467, 1903.

additional oxygen atoms needed in each unit cell of magnetite are indicated by four dotted circles in Fig. 1. These positions, however, are too close to the new ferric atoms, while the old O atoms are now too far away from the ferric atoms. A general readjustment of equilibrium becomes necessary, resulting in the distortion of the cube lattice to a rhombohedral one with entirely new elements of symmetry.

It is known that gases diffuse in metals. This process is usually not a chemical reaction. However, the diffusion of solids in solids in isomorphous mixtures (solid solutions) is in the nature of replacement of one atom by another and might be grouped under chemical reactions. In the case of oxidation of magnetite the objection might be raised to classification of oxidation under diffusions, that a new space lattice and a new compound is formed. On the other hand, the process of oxidation is relatively slow and proceeds from the outside of grains or planes. The O atoms advance step by step from one new position to another. We can not assume that O atoms can "push" through a completely occupied space lattice of hematite for example. Langmuir [18] proposes a good working hypothesis on the progress of atoms in solids.

Inspection of the positions of the hypothetically added O atoms in Fig. 1 reveals the fact that they are on the octahedral planes, which have the weakest bonds and give rise to parting or cleavage. It is easy now to understand why oxidation should proceed along these planes. One might argue that these planes are so closely spaced that the total visible effect of oxidation in these planes should be the same as if oxidation proceeded uniformly over the entire surface of the magnetite grains regardless of orientation. This argument would not be valid in view of the fact that phenomena like etch figures, striations, solution cavities, and others do not effect all of the equivalent atoms in a surface equally and all at the same time.

It is probable that the familiar replacement structures along

[18] Jeffries, Zay, *op. cit.*, p. 7. Also reproduced by E. E. Fairbanks, ECON. GEOL., vol. 20, pp. 470–484, 1925.

well defined crystal planes (Schneiderhöhn's [19] Entmischlungsstruktur), in many minerals can be explained by similar reasoning. The well known lamellae of ilmenite ($FeTiO_3$) in magnetite [20] are parallel to the octahedral planes. The segregation of ilmenite from a solid solution of magnetite and ilmenite takes place at high temperature. Before this temperature is reached in cooling, the Ti atoms are probably evenly distributed through the lattice.[21] When the temperature of segregation is reached the Ti atoms (including the necessary O) become unstable in their positions and naturally shift to planes which will have the weakest bonds with respect to magnetite. It seems very probable that the space lattices of hematite and ilmenite have definite orientations with regard to the space lattice of magnetite. The writer believes that such orientation is brought about by the sharing of atoms common to both solids along planes in which the spacing of these atoms is approximately alike, or at least similar, in both lattices.

A CURIOUS CASE OF OXIDATION.

When specimen M 709 (see table) was examined before heating it consisted of magnetite entirely (except gangue). When reexamined after three days of heating at 200° C., hematite had begun to form as minute rods along octahedral planes. Close inspection revealed that these rods in some places had arranged themselves in large former crystals of specular hematite (Plate I, C, D). The arrangement of the little rods might be compared to a "cross stitch design" outlining the specular crystals.

The author once before [22] has shown that primary hematite can be replaced by magnetite and again change back to hematite. Gilbert [23] seems to think that the hematite which I called hematite of the second generation really antedates the magnetite in this

[19] Schneiderhöhn, H., "Segregation Phenomena of Ore-forming Solid Solutions, etc.," ECON. GEOL., vol. 19, pp. 298–300, 1924.

[20] See, for example, Singewald, J. T., Jr., "The Microstructure of Titaniferous Magnetites," ECON. GEOL., vol. 8, pp. 207–214, 1913.

[21] Tamann, G., "Zur Atomistik chemischer Reaktionen in anisotropen Stoffen," *Z anorg. Chem.*, vol. 26, pp. 263–276, 1915.

[22] *Op. cit.*

[23] Gilbert, G., *op. cit.*, p. 591.

particular case. This view would not explain the sawtooth edge (Plate II. of my original paper) of the crystals partly altered back to hematite.

In Plate I. C, D, it will be noticed that the "cross stitch pattern" cuts euhedral grains of magnetite. This coming back of the original outlines of replaced grains after millions of years may be explained in two ways. When hematite changes to magnetite Fe must enter the lattice with an increase in volume, or O must leave the hematite with a corresponding decrease in volume. This latter process is the more likely one. This diffusion of O atoms toward the surface would cause the formation of magnetite from the centers of the grains outward, explaining Gilbert's observations.[24] It is possible that practically the required number of O atoms leave the lattice. If, however, minute submicroscopic amounts of the hematite remained scattered through the newly formed magnetite, reoxidation to hematite probably would start with these minute nuclei as beginnings. On the other hand, all of the O atoms in excess of the number needed for magnetite might leave the crystal. The lattice, however, instead of becoming perfectly cubic would remain slightly rhombohedral on account of certain straining effects. Such strain effects might be so slight that they could not be detected by X-rays,[25] yet they would cause preferred oxidation of these slightly distorted areas.

DO NATURAL SOLUTIONS OF MAGNETITE AND HEMATITE EXIST?

Sosman and Hostetter[26] believed that these two minerals are almost completely soluble in each other. Broderick[27] has crititized their investigation. The present writer[28] expressed his belief some time ago that solid solutions are probably not found in nature. It is true that *artificial* Fe_3O_4 and Fe_2O_3 show a num-

[24] *Op. cit.*, p. 591.

[25] Rinne, F., "Feinbauliche Erörterungen und röntgenographische Erfahrungen über optische Anomalien," *Centralbl. Min.*, 1925, pp. 225–241.

[26] Sosman, R. B., "The Ferrous Iron Content and Magnetic Susceptibility of Some Artificial and Natural Oxides of Iron," *Trans. Am. Inst. Min. Eng.*, vol. 58, p. 433, 1918.

[27] Broderick, T. M., *op. cit.*

[28] *Op. cit.*, p. 3.

ber of peculiarities, for instance the retention of magnetic susceptibility in Fe_2O_3 oxidized from Fe_3O_4 in a stream of oxygen at 220° C. Welo and Baudisch [29] even claim that such Fe_2O_3 retains the lattice of magnetite up to 550° C. In natural minerals such properties are either absent or can be explained by inclusions of one mineral in the other.

To be certain with regard to the structure of martite, X-ray diffraction patterns [30] were made of powders of: martite, blue ore from Mahoning Pit, Hibbing, Mesabi range; martite, Twin Peaks, Utah; specular hematite, Isle of Elba; magnetite, Port Henry, New York; goethite crystals, locality unknown.

The two martites and the specular hematite gave absolutely identical patterns. Magnetite and goethite are quite different from martite.

Solid solutions on reaching a temperature of instability break up into their components, and provided the drop in temperature is slow enough for the atoms to move relatively large distances, this segregation becomes visible under the microscope. In many solid solutions the solubility gradually diminishes and a continually varying composition of the solid is the result. These changes from the centers of grains toward the margins give rise to so-called zones in some solids. (Coring of metallographers.) It seems that in geological processes there should *always* be enough time for solid solutions to break up into their components (except in very small igneous masses), if the solid solutions are not stable at low temperatures. There is no doubt that most of the deposits mentioned in the table did not undergo much change in temperature and pressure if geologically short periods of time, like a few years, are considered. In other words there was plenty of time for the segregation of the solid solutions at any stage when they

[29] Welo, L. A., and Baudisch, O., "The Two-stage Transformation of Magnetite into Hematite," *Phys. Review*, vol. 25, p. 587, 1925. See also: *Science*, vol. 62, p. 311, 1925. The paper "Ferromagnetic Ferric Oxide, Artificial and Natural," by R. B. Sosman and E. Posnjak (Jour. Wash. Acad. Sci., vol. 15, p. 329, 1925) has just come to my attention. While very important it does not affect the present problem materially.

[30] The writer wishes to acknowledge the help of Dr. O. E. Harder in taking the X-ray spectograms.

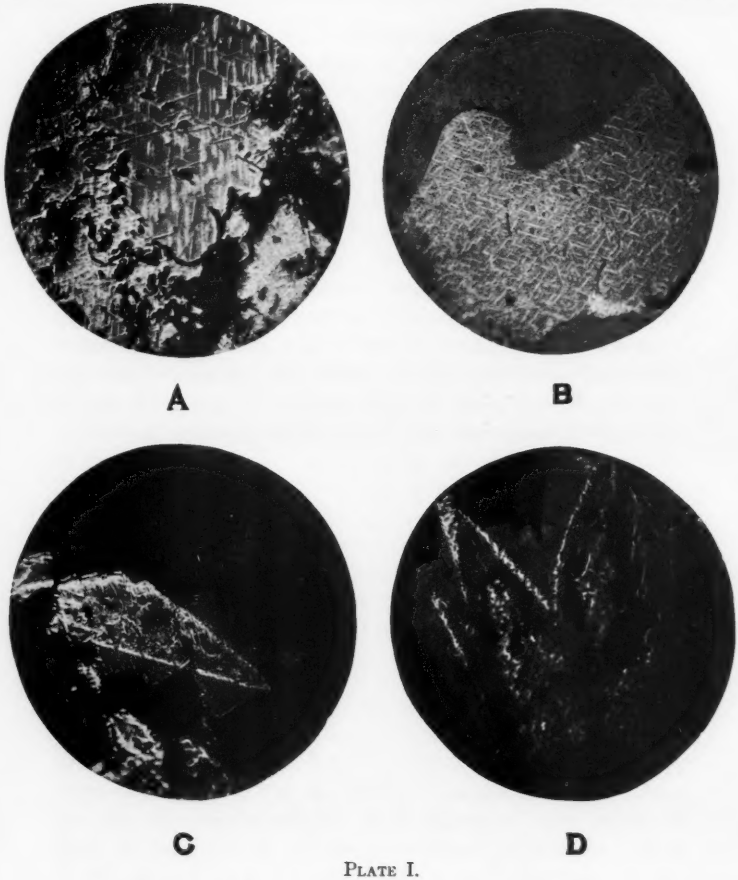

PLATE I.

(*a*) Soudan magnetite (gray) oxidizing along octahedral planes to hematite (white).

(*b*) Titaniferous magnetite oxidizing along octahedral planes. Ilmenite occupies the large uniformly gray areas.

(*c*) Outline of an early specular hematite crystal (white) brought out in magnetite (light gray) by artificial oxidation.

(*d*) A group of early hematite crystals brought out as in (*c*).

reached instability. It is a well established fact now, that practically all magnetites regardless of origin have microscopically visible hematite in them. With respect to magnetites formed at high temperatures this would mean that the minreal was saturated with hematite a short time before its atoms reached a state where diffusion became impossible. Otherwise hematite would not have segregated. The possibility of zonal gradation from pure magnetite to pure hematite is excluded, because no polished specimens show it. The contacts of the two minerals always show sharp color contrasts. It is obviously impossible that all magnetite containing inclusions of hematite is a solid solution. The same reasoning applies even more forcefully to magnetites in sedimentary iron formations. Let us suppose that a solid solution existed a little above 200° C., but was unstable below that temperature. Visible segregation probably would have taken place in the ovens in 330 days at 200° C. The hematite, however, that formed in that time is clearly due to artificial oxidation. Otherwise it would not be confined to a thin layer on the surface of most specimens, nor would it diminish from the surface inward. Another proof of insolubility at temperatures up to 200° C. are the very sharp lines of demarcation between the two oxides in the treated specimens.

AGE RELATIONS OF THE OXIDES.

In many specimens, especially in the Soudan formation of the Vermilion range, almost perfect euhedra of specular hematite and octahedra of magnetite occur independent of one another in the same specimen, as if they had formed contemporaneously. There are also instances in which a grain appears to have started to grow as magnetite but suddenly continued as hematite. Some of the specular crystals of Elba seem to be of this type.

Besides the octahedral type of replacement there is a replacement of magnetite which follows no definite planes. This is very common and can be observed at all stages. In many cases it is impossible, therefore, to tell whether the magnetite or the hematite is the older. In polished specimens oxidation was most intense

right on the surface, though following definite planes. It is quite probable, judging by the very shallow penetration of oxidation in most specimens that the minute rods of hematite will become thicker as shown in Pl. I. *A*. and grow together presenting a "solid front" before penetrating to much greater depth. This would result in the type of replacement just mentioned.

Why some magnetites oxidize much more rapidly than others was explained partly by difference in grain size. There must be still another reason due to conditions of origin since the magnetites formed at high temperature seem to be the more resistant ones, (Daiquiri excepted). Two possible reasons are: (1) that a β modification of magnetite exists [31] above 530° C. and the α modification derived from it retains some properties not possessed by relatively low temperature magnetites, (2) that magnetite derived from limonite and hematite has a slightly distorted space lattice as previously suggested.

SUMMARY.

Martite, regardless of origin, has the same internal structure as specular hematite.

Oxidation of magnetite in air at 150° C. and 200° C. always begins along octehedral planes. An explanation for this fact as well as for the cleavage or parting is found in the arrangement of the atoms in the space lattice. The common octahedral form of magnetite and the tabular habit of specular hematite (especially micaceous) are also explained by the space lattices.

In the oxidation of magnetite to hematite by addition of oxygen an increase of volume of 5.2 per cent takes place instead of 2.5 per cent, as formerly believed.

Magnetite in air at 150° C. and 200° C. oxidizes extremely slowly. The rate depends on the size of the grains and on the origin of the magnetite. It may be hundreds of times faster in one specimen than in another, which should explain why unaltered magnetite, martite and specular hematite occur under so many apparently incompatible conditions in nature. The large free surface energy in very small grains (apart from the fact that their

[31] Königsberger, J., and Schilling, K., *op. cit.*

total exposed surface is very much larger) is the reason that they oxidize much more rapidly. This free energy seems to be reduced by oxidation of the magnetite. Why the origin of magnetite affects its rate of oxidation may be explained either by derivation from a high temperature β modification, or by a slight distortion of the lattice of magnetite of sedimentary origin.

No solid solutions of magnetite and hematite seem to exist in nature.

When hematite is reduced to magnetite by removal of oxygen, magnetite should grow from the inside of hematite grains outward. A curious case of oxidation shows that in magnetite which was partly specular hematite at one time, the outlines of the specular crystals can be brought back by artificial oxidation.

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EDITORIAL

REPLACEMENT COPPER ORE DEPOSITS IN QUARTZITE.

REPLACEMENT deposits of sulphide ores in siliceous sedimentary rocks are rare. Therefore, the occurrence of large massive deposits of high-grade copper ore in quartzite, recently discovered and now being extensively worked in the Ashio mine will be of some interest to economic geologists.

The Ashio mine in the province of Shimotsuke has been famous as a producer of copper for several hundred years and it is to-day the largest mine in Japan in which copper veins are worked. The region that includes this mine is composed of Upper Paleozoic formations consisting of clay slates, quartzites, and subordinate limestones, intruded by a liparite mass that is a volcanic neck and has a nearly circular outcrop with a diameter of about two miles. More than two hundred veins are found in this district, most of which are enclosed in the liparite, though a few of them are in clay slate or cut both the sedimentary and intrusive rocks. They are divided into two parallel vein systems; one strikes northeasterly, and the other westerly. The two groups often intersect intricately. The ore deposition is evidently related to the liparitic eruption of late Tertiary age. The veins consist chiefly of quartz, chalcopyrite, and pyrite with varying amounts of pyrrhotite, zincblende, galena, rarely associated with fluorspar, apatite, wolframite, or vivianite. In and below the oxidation zone, cuprite, malachite, azurite, native copper, bornite, chalcocite and other minerals are extensively developed.

Besides the filled fissures, many irregular massive deposits of replacement origin, called "kajika" by miners, have been known to exist in the liparite, in close relation to the veins. They are

usually found at the intersections of small fissures that probably served as the passageways for hydrothermal solutions. Irregular pipe-shaped ore masses have been formed, surrounded by intensely sericitized liparite and often connected by means of small veins with larger veins. The ore of the "kajika" deposits is characterized by the scarcity of gangue minerals and by the presence of high-grade massive chalcopyrite.

It has long been believed by the miners of this property that neither rich veins nor replacement deposits occur in quartzite. A few years ago, however, a massive deposit or a "kajika" was discovered by chance in quartzite, and by later prospecting numerous other ore masses of the same character have been found. At present they are the chief sources of high-grade ores for the Ashio smelters.

The wall rock of the newly found "kajika" deposits is cherty quartzite, a characteristic member of the Japanese Paleozoic formation, which is composed of cryptocrystalline silica, and in some places is recrystallised into interlocking grains of quartz. The deposits are rudely tabular or irregularly massive in form, and usually pitch downward, sometimes attaining a length of 100 feet or more. They consist of massive chalcopyrite mixed with more or less pyrite, pyrrhotite, and other sulphides, and replace the wall rock with sharply-defined but intricately embayed boundaries. In some places, especially along the margins of the ore bodies, the massive ore passes into veinlets or a network of veinlets, filling joints, bedding planes, and other cracks, and containing vugs lined with crystals of pyrite, chalcopyrite, quartz, and apatite. It is noteworthy that these veinlets become more and more quartzose toward the outside and finally thin out. Corroded fragments of quartzite, cemented by chalcopyrite and other sulphides are abundantly present near the border of the ore masses. In general, these replacement deposits occur in places where cracks abound, large ore bodies being found near synclines where tension cracks were richly formed. Ore masses formed along fault-breccias are also known. In all cases, replacement of quartzite by sulphides is evident.

The "kajika" deposits of the Ashio mine formed in quartzite are instructive, because they show that replacement advances steadily even in highly quartzose rocks where they are exceedingly shattered and when mineralizing solutions are strongly alkaline. In this connection I recollect the famous mercury deposits of the Almaden mine in Spain, where Silurian quartzite strongly impregnated with and almost wholly replaced by cinnabar forms such important ores.

TAKEO KATO.

DISCUSSION AND INFORMAL COMMUNICATIONS

RELATION OF EARTH TEMPERATURES TO BURIED HILLS AND ANTICLINAL FOLDS.

Sir: This paper, by W. T. Thom (ECON. GEOL., vol. 20, Sept.–Oct., 1925, pp. 524–530), based on observations of underground temperature by C. E. Van Orstrand, brings out the close relation between Wyoming anticlines and heat gradients. The rock temperature at any depth under the crest of Salt Creek, Warm Springs, Ferris and Lost Soldier domes is higher than in the surrounding country. Thom regards this as a relic of the distribution of temperature resulting from the Eocene deformation. The underground isogeothermal surfaces doubtless were approximately horizontal before the Eocene deformation which on anticlines would lift the isogeotherms along with the strata. Could so early an uplift of isogeotherms persist until this day? The reviewer does not know, but he is surprised by the idea, which should be criticized by a geophysicist capable of calculating how much of such differences of temperature can last through scores of millions of years.

In his Fig. 2, Thom shows that at a depth of about 300 feet the isogeothermal surface of 60° F. is nearly level, or nearly parallel to the general surface of the ground, but that the deeper isogeotherms are folded along the same axis as the strata but at a lower angle. The isogeotherms of 80°, 100° and 120° F. are drawn as having been lowered along the axis about 600 feet more than on the limbs, a difference assigned to radiation of heat after the folding. These three lower isogeotherms are nearly parallel, indicating nearly uniform upflow of heat across them, and indicating accelerated radiation along the axis above the isogeotherm of

80° F., apparently within some 600 or 800 feet of the ground surface. It would seem from the diagram that heat radiation has been unable to destroy the fold of the isogeothermal surfaces below this depth, nor to do more than lower the fold of the isogeotherms by less than half its original height. Meanwhile one or two miles of strata have been removed by erosion, requiring corresponding descent of all the isogeotherms. Could they retain half their Eocene fold while descending one or two miles? Could a 2,000 ft. arch of the isogeotherms retain nearly half its height, while the isogeotherms were being slowly lowered through one or two miles of rock?

The arching of the isogeotherms is about as strong at a depth of 1,000 feet as it is at 3,000 feet. Between these depths lies the First Wall Creek sand, in which there is reason to believe there is a relatively strong current of water flowing northwest, towards the lowest outcrop of the sand on Powder River. This current came from the south, where the sand has its highest outcrop on the flank of Casper Mountain. Indications of the northward element of this flow are found in the northward tilting of the water level below the oil in the First Wall Creek sand, and in the dilution of the water[1] with the addition of carbonates as compared with the water in the Second Wall Creek sand, where circulation was slow on account of the finer grain of the sand.

The circulation of water in the thick porous Dakota, Sundance and Tensleep sands must be much greater than in the Wall Creek sands, and the most rapid flow must follow the same direction from highest to lowest nearby outcrop, moving northwestward near Salt Creek. This circulation carries the water of all the sands down under a syncline that is many thousand feet deeper than the top of Salt Creek dome. In rising from the syncline the water must carry heat toward the crest of the anticline, warming the crest above its normal temperature for any depth below about 500 feet. The circulation of water therefore acts like the heating system of a house which lifts hot water from a furnace in the basement.

[1] Cf. E. L. Estabrook, *Bull. Amer. Assoc. Petrol. Geol.*, vol. 9, No. 2, p. 239, 1925.

Thom dismisses this idea because he seems to think that the water in the Wall Creek sands would enter at the nearest outcrops along Powder River valley west of Salt Creek and flow eastward down the dip. An eastward direction of flow is improbable because the Wall Creek sands thin out eastward disappearing entirely under a basin of Fort Union strata, east of which they never are seen. It is also improbable because it does not correspond with our observations of the directions of flow of water in sands at other places, as in the Bighorn Basin where cumulative evidence points toward a generally active flow along the periphery of the basin, from highest to lowest nearby outcrop, with relative stagnation toward the center of the basin. The Powder River outcrops west of Salt Creek are the lowest in the region and therefore are not intakes but are outlets of water. E. L. Estabrook and others are preparing a paper giving further details about the lines of major circulation in Wyoming oil sands and their influence on the accumulation of oil and gas.

In all the four Wyoming domes mentioned by Thom the direction of water circulation in the sands is thought to take it under deep synclines before it reaches the domes. In addition to this the great flows of hot water near Thermopolis suggest that the Warm Springs dome may have some other source of hot water to heat it. The circulation of water under Big Muddy field does not cross a deep syncline, and a test of its underground temperatures would help decide between the rival hypotheses.

For the time being the reviewer considers transfer of heat by water circulating in sands a most probable explanation of the abnormally high temperatures under Wyoming oil fields.

CHESTER W. WASHBURNE.

GISBORNE,
NEW ZEALAND.

ELECTRICAL CONDUCTIVITY OF ORE-MINERALS.

Sir: Mineralogists owe thanks to Messrs. Kerr and Cabeen for the contribution which they have made to our methods of identification in their recent article under the above title (ECON.

GEOL., vol. 20, Dec., 1925, pp. 729–737). It happens that recent experiments which the present writer has been conducting during spare moments overlap those of Kerr and Cabeen particularly in connection with their class 3, "non-conductors," in their table of comparative electrical conductivity of ore-minerals (p. 735). It is possible by simple means to obtain quantitative data of the comparative insulating power of these minerals listed as non-conductors. The method of determining insulating power, or dielectric constant, not only has the advantage of being simple, even more so than this new method of determining electrical conductivity, but it is a very sensitive method, whereas the methods employed by the physicist for delicate electrical conductivity measurements are tedious and involved, as the authors of the above paper state.

If single-phase alternating current is available, giving a range of from say 100 to 340 cycles, permitting the voltage to be altered from 100 to about 400, all that is necessary to determine dielectric constants are various liquids of different dielectric constants, a small watch glass, two needles set in a penholder so that there is a small gap between the points and a binocular microscope for observing the attraction or repulsion of the mineral powder.[1]

The English investigators do not give methods of determining dielectric constants beyond the range from 2 to 36. This range will catch many of the transparent minerals but petrographic methods are far superior aids in the identification of minerals that transmit light. The opaque minerals, however, may offer many difficulties in identification and any quantitative data which possess diagnostic value are most welcome.

In the preliminary experiments performed by the writer, pure distilled water with a dielectric constant of 81 was the highest liquid used, although other liquids possessing higher constants are known. The media must be chosen, however, to avoid electrolytic effects. With water as a medium it is possible to dis-

[1] For further information on methods see "Dielectric Mineral Separation: Notes on Laboratory Work," by B. W. Holman, *Bull. Inst. Min. and Met.* (London), Feb., 1924, pp. 1–7.

tinguish between very high constants and those greater than 81 although not far from that value. If the dielectric constant of the mineral is very much greater than 81 the particles will bridge the gap and form an arc. In the case of minerals greater than 81 but not too far from that value, the particles may bridge the gap but never produce an arc.

The following results of the preliminary experiments indicate the possibilities of the new method:

Boulangerite	36	Polybasite	>81 arc
Carnotite	9.3	Pyrargyrite	>81
Cuprodescloizite .	7.7–8.6	Rutile	>81
Franklinite	>81	Scheelite	7.7
Jamesonite	>81	Stephanite	<81 near
Limonite[2]	9.9–15	Stibnite	>81
Miargyrite	81	Vanadinite	10.8–11.2
Monazite	8.2	Wulfenite	11.7

It is important to emphasize the two following considerations. First, that the dielectric constant as well as other physical properties varies with direction in anisotropic media; thus the c axis of andalusite is the direction of greatest light velocity and also the direction of the highest value for the dielectric constant. In the case of andalusite from the Tyrol this value amounts to 12. It may be necessary to alter the voltage in order to minimize "electric wind" effects. If this should prove necessary the voltage should not be made too low, otherwise the attraction or repulsion will not be observed.

With more experimental data forthcoming, especially with regards to constants above 81, this new method should prove of great value to the mineralogist in conjunction with determinations of electrical conductivity. The economic geologist should not lose sight of the fact that problems of genesis, zoning, and many others involve accurate mineral determinations.

ERNEST E. FAIRBANKS.

RENO, NEVADA.

[2] The limonite sample given above consisted of two constituents, one giving the constant 9.9 and the other the value 15. Note that rutile does not arc in water of dielectric constant of 81. The value of the powder is listed as 110.

ANGULAR INCLUSIONS IN ORE DEPOSITS.

Sir: In your December, 1925, number, Mr. J. E. Spurr, in a discussion of a previous paper (Oct., 1924) by Prof. Alan Bateman, and a discussion of it by C. A. Merritt (Sept.–Oct., 1925), took up the method of the floating of rock inclusions now found in veins, and made a statement which, if correct, is of great geological importance.

He states that a magma at a high temperature would be almost as fluid as water, and therefore to have magmatic colloidal jelly-like material of a sufficiently glutinous consistency to buoy up rock fragments as seen in many veins, the colloidal gelatinous material must have been comparatively cool at the time of its injection into the fissures where it is now found as solid vein- and ore-matter.

How to reconcile the seeming inconsistency of magmatically injected material, with a degree of coolness such as that postulated, appears to be difficult, and would seem to require some special quality in the colloidal material whereby its heat-carrying capacity, compared with the material with which it was associated at depth, was greatly reduced.

I have never seen anything on this point in the geological literature, but an observation that attracted my attention would seem to bear out the possibility of a lack of heat-carrying capacity for colloidal material such as that under discussion.

A tank of oil, part of which was sump-oil highly charged with an emulsion of oil and water was heated, in order to break down the emulsion, at a temperature of 210 degrees Fahrenheit, and held there for several hours. The tank was then allowed to stand for two or three days to permit the water to settle. When this had been accomplished, it was noticed that the side plates of the tank in contact with the settled water were still hot, as were also the plates higher up in contact with the oil, while between the two there was a band of the plates about a foot wide which was distinctly cold to the touch, and which band later proved to have been in contact with a layer of emulsion between the oil and the

water; emulsion that had resisted the heating process for breaking it down into oil and water.

No temperature records were made at the time, but as the plates in contact with the emulsion were cold to the touch, while those in contact with the water and oil were quite warm, and the water as it was drawn off was unbearable to the hand, a wide range of temperature is shown. This is the more remarkable when it is considered that the cold emulsoidal material was interstratified between the hot water and the hot oil.

The general characteristics of a molecule of emulsion formed of oil and water, are no doubt not greatly dissimilar to a molecule of colloidal silicic acid, which too, like the oil emulsion varies in its degree of hydration, and if the principles underlying the cause of the oil emulsion resisting impregnation by heat are applicable to magmatic colloidal material, then it would be possible for colloidal vein- and ore-forming material to exist at depth in a comparatively cool state in an environment of a much higher temperature.

The evidence of comparatively low temperatures during their formation, which is presented by deposits of vein- and ore-minerals filling fissures and shattered zones, would seem to substantiate this possibility.

JACOB W. YOUNG.

MONTEBELLO, CALIFORNIA.

REVIEWS

Versuch einer natüralichen Klassification der im weiteren Sinne magmatischen Erzlagerstätten. By PAUL NIGGLI. 69 pages. Wilhelm Knapp, Halle, 1925.

This paper is the first of a series on economic geology promised us by the editor, Dr. Georg Berg. Niggli, as in all his books, emphasizes the inseparable nature of igneous rock provinces and mineral deposits. He starts with a discussion of the physical and chemical side of the question, leading up to the all pervading thought that every ore deposit of magmatic origin is really a magmatic differentiate. In this respect he agrees closely with J. E. Spurr without, however, mentioning the probability of vein dikes except in special cases. His paragenesis of the chemical elements and the development of the order in which the minerals come out of the magmatic differentiates from the liquid magmatic to the hydrothermal stages are illustrated by clear and concise tables.

Niggli's systematic classification of deposits is the following. Lindgren's [1] corresponding terms are given in parentheses.

I. PLUTONIC.

A. Hydrothermal (mesothermal and part hypothermal).
B. Pegmatitic-pneumatolytic (pyrometasomatic; exudation veins, deep-seated type; pneumotectic; and part of hypothermal).
C. Liquid-magmatic (orthotectic).

II. VOLCANIC.

A. Exhalent to hydrothermal (epithermal; sublimates; in part exudation veins of surface type).
B. Pneumatolytic (in part exudation veins of surface type).
C. Liquid-magmatic.

This table is based primarily upon geologic-physical considerations and only secondarily upon chemical-mineralogical ones. Niggli takes up each division somewhat in detail. He introduces a more or less new subgroup when he puts certain Lake Superior iron ores and similar deposits under the heading "volcanic-subaqueous and simultaneously biochemical."

In the last part of the paper the distribution of deposits with respect to

[1] ECON. GEOL., vol. 17, pp. 292–294, 1922.

age and position is taken up. The great pre-Cambrian shields surrounded by the Paleozoic bow areas and encircled by the Mesozoic and Cenozoic folds and uplifts are discussed.

In connection with the kinds of deposits associated with the three rock divisions, basic, intermediate, and acid, Niggli discusses the following question: Considering all deposits as differentiates, at what stage of the primary segregation of the original gabbro-basalt magma into basic and acidic masses is it possible and probable for the various deposits to form? Fe, Mn, Pt, Cr, Ti, Ni, Co, P, Cu, (to some extent Ag, Pb, Zn, Au, Bi) deposits will form entirely or may form in the earlier basic magmatic stage. Sn, Mo, W, as well as gold deposits proper, do not form till the liquid magmatic differentiations have concentrated these elements considerably in the more acid magmas. Between these two extremes occur Pb-Zn-Ag and Cu deposits in considerable numbers, evidently accompanying the differentiates leading to the granodioritic and granitic magmas, though leaving them before the true granitic composition is reached.

Such deposits as the Pb-Zn deposits of the Mississippi Valley, of the Rhein and Belgium, of Upper Silesia, and of the Austrian Alps are called "telemagmatic." They are not thought to be connected with any small intrusions close to the deposits (such intrusions are very rare anyway) but with distant basic magmas intruded during two successive periods of folding. Some Ni-Co-Ag veins (Upper Erzgebirge) are also thought to be telemagmatic.

It may be mentioned that among the large number of books and papers published by Niggli in the last few years, the book entitled "Die leichtflüchtigen Bestandteile im Magma" (Leipzig, 1920), dealing especially with ore deposits, is practically unknown in this country. Its first edition is sold out, but a second one is promised by the author.

JOHN W. GRUNER.

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MINNEAPOLIS, MINN.

Geology and Mineral Resources of the Philippine Islands. By WARREN DU PRE SMITH, Manila.

This work is an attempt to collect into one volume a summary of all information of interest to geologists, mining engineers, students, and mining investors. The author has succeeded admirably in presenting a vast fund of interesting and useful information about the Philippines. His style is much clearer than that of most geologists. Although the material is compressed, the writing is not disjointed. These qualities calculate to make the work of interest to the casual reader as well as those seeking information.

The chapter on geography outlines the general position of the islands, describes the climate, flora, and fauna. There follows an instructive chapter on physiography which contains a section dealing with the human response to physiographic conditions. The chapters on Stratigraphy and Correlation, Petrography, Paleontology, and Mineralogy are complete and contain many charts. There are interesting chapters on Seismology and Volcanology. A long chapter is devoted to Regional Geology, in which each island is treated according to its importance and points of interest. The chapter on Economic Geology contains descriptions of the important coal, oil, gold, and iron mining fields; deals slightly with the methods of mining, milling, and smelting. Other mineral resources are treated as their importance warrants.

The mineral products in order of their economic value are gold, salt, stone, coal, sand and gravel, lime, clay, iron ore, mineral waters, bituminous rock, silver, and sulphur.

The appendix contains an abstract of the mining laws of the islands, a directory of mine operators, and tables of mining costs. In the set of 39 plates are interesting views of the country, fossils, thin sections, and several maps, including a hydrographic and relief map of the islands.

C. R. M. TUTTLE.

Who's Who among North American Authors. Vol. II., 472 pp. Golden Syndicate Publishing Co., Los Angeles, 1925–26. Price $5.00.

This volume contains biographical and literary data of living authors and a section devoted to editors, magazine writers and press editors. The biographies are arranged alphabetically and are clearly typed and concise. Individual names may readily be found. A useful part of the supplement is an index of authors by states and a list of pseudonyms, with their authors' names.

The book is indispensable to those who desire to refer to authors and their works.

A. M. B.

Climates of the United States. By ROBERT D. WARD. 518 pp; 144 figs. Ginn & Co., Boston, 1925. Price $4.00.

This book is the outcome of twenty-five years of teaching experience at Harvard University, and incorporates much material previously published by Professor Ward in scientific journals. It deals exhaustively with climates and weather and their controls and causes. Two chapters also consider climate and health, and climate and crops. Those who desire information about climate and weather need not go outside its pages, and abundant footnotes give the sources of material.

It is a book primarily for the student of climate and weather, though the lay reader may derive much of value from it. The latter's interest, however, is apt to be dampened by the mass of detail it contains. The writer's style indicates the scientific, not the popular writer. The book is up to date, clearly written, well typed and bound, and appears to have had careful proof reading.

A. M. B.

Economic Geology of the Ayrshire Coalfields, Area I. By J. E. Richey, G. V. Wilson and E. M. Anderson. **Area II.** By E. M. Anderson and G. V. Wilson. Geol. Surv. of Scotland, Edinburgh, 1925. Price, 2/6 and 3/; Map 2/6.

The former of these two memoirs is the first of a series to be devoted to a description of the Ayrshire coalfields. A short summary of the geology of the district, including Kilbirnie, Dalry and Kilmaurs, is followed by a detailed account of the formations. These comprise the Carboniferous Limestone series, the coals, ironstones and limestones of which have been long and extensively worked; the Millstone Grit with its valuable fireclays and bauxitic clay; and the Coal Measures of the Saltcoats, Stevenston and Kilwinning area. In the final chapter non-metallic materials are described, and tables are given of analyses of clays, limestones, sandstones and ironstones.

Area II. includes the greater part of North Ayrshire, and deals mostly with the Kilmarnock Basin Coal Measures. Most of the Coal Measure coals have now been extracted; the coals in the Limestone Coal group are mainly intact and have been reached in several instances by boring. They are, however, fewer than in the corresponding group in Area I., and cannot be regarded as forming any very large reserve.

A geological sheet of vertical sections of the North Ayrshire Coalfields accompanies the memoirs.

A. M. B.

SCIENTIFIC NOTES AND NEWS

V. A. Gilles is at Billings, Montana, acting as assistant chief geologist for the Northern Pacific Railroad Company.

T. B. Nolan, of the U. S. Geological Survey, is continuing field work in the Gold Hill District, Utah.

Frank B. Adams of Montreal is to attend the Pan-Pacific Congress in Tokyo, Japan, next October.

Charles Hodge Scott, oil geologist, died at Princeton, N. J., on March 21. He was a son of Professor William B. Scott and a member of the class of 1909 at Princeton University.

H. T. Stearns, of the U. S. Geological Survey, is in charge of a ground-water investigation in the Mokelumne River Basin of California.

Howland Bancroft of Denver, Colorado, is to sail soon for Venezuela, having been elected vice-president and general manager of the Lago Petroleum Corporation.

W. R. Brown is in Chile as assistant general manager of the Anglo-Chilean Nitrate Corporation.

Arthur C. Terrill, who has been connected with the Pei-Yang University at Tientsin, China, is now petroleum engineer at the State Mining Bureau at Los Angeles, California.

George C. Branner, state geologist of Arkansas, is making a geological survey of the coastal-plain area of Arkansas, which is to cover about half the state.

Leo Ranney is president of the Ranney Oil Mining Company, a company formed to take up technical problems of engineering and geology in the recovery of oil by mining operations, with offices in New York City.

C. E. Julihn, of the U. S. Bureau of Mines, is now official engineer in charge of copper investigations.

The American Association of Petroleum Geologists has elected the following officers for next year: President, A. W. McCoy; Vice-President, C. R. McCollom; Secretary-Treasurer, Fritz Aurin; Editor, J. L. Rich.